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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PALEOZOIC PALEOGEOGRAPHY OF ARIZONA

BY ALEXANDER STOYANOW

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ABSTRACT

In Arizona two major areas of deposition separated by Mazatzal Land had already been established by late pre-Cambrian time. The first Paleozoic transgression from the northwest, starting possibly late in the Early Cambrian and depositing the Tapeats sandstone, reached central Arizona. The southern limit of the Bright Angel shale and the overlap of the Muav limestone on the Tapeats indicate subsequent movements northward and southward in early Middle Cambrian time before the final withdrawal of the Cambrian sea from northern Arizona. In southeastern Arizona, likewise, only the earliest phase of the Cambrian transgression reached the central part of Mazatzal Land; the northwestern boundaries of the late Middle Cambrian Troy, Santa Catalina, Cochise, and the Upper Cambrian Abrigo formation are all successively regressive southeastward. Ordovician strata are known in Arizona only in the areas near New Mexico. Silurian and early Devonian deposits are present only near the extreme peripheral margins of Mazatzal Land beyond the limits of the State. Central-western Arizona was a positive element during the Ordovician, Silurian, and both the Early and Middle Devonian. The overlapping late Devonian transgression, also initiated from the northwest and southeast, deposited clastics around the central part of Mazatzal Land and was succeeded by deeper sea conditions in Mississippian time. The southeastern trough was the deeper

of the two basins throughout the Late Paleozoic and is regarded as an arm of the Ouachita seaway. Following the retreat of seas from most of Arizona in the Late Mississippian a marked early Pennsylvanian transgression was introduced from the southeast and limited on the northwest by Mazatzal Land. Grading of the Manzano gypsiferous series and fossiliferous limestones into the Supai northward and distribution of Kaibab faunas establish the relation of the southeastern trough to the northern area of Permian shallow marine and continental deposition.

INTRODUCTION AND ACKNOWLEDGMENTS

The writer (Stoyanow, 1936) has briefly discussed the Paleozoic waterways of Arizona; to outline these seas required many additional observations. Dr. G. M. Butler, former Director of the Arizona Bureau of Mines, enabled the writer to study, in connection with the summer programs of geological work of the Bureau, some of the most critical areas. Grateful acknowledgments are also due Dr. Eldred W. Wilson, Geologist of the Bureau, who willingly helped both in the field and in the office, and Mr. L. F. Brady, who assisted the writer on many of his trips. This paper discusses the distribution of Paleozoic basins and lands in Arizona on the basement complex and their relation to the main neighboring seaways and land masses. The accompanying maps are based on factual outcrops and distribution of strata. The imperfections of many of the postulated views are fully recognized. It is hoped that the facts presented will contribute toward the interpretation of some of the major paleogeographic problems of the Southwest, and a critical evaluation of suggested solutions will be greatly appreciated.

BASEMENT COMPLEX OF ARIZONA

Ransome (1916, p. 133) separated Arizona topographically into three principal regions: the plateau region which occupies the entire northern and northeastern parts of the State; the mountain region which extends from southeastern Arizona northwestward beyond the central part of the State, roughly paralleling the southwestern margin of the plateau region; and the desert region of southwestern Arizona. For paleogeographic studies, Arizona is readily separated into two regions: (1) the plateau on the northeast, where the thick, nearly horizontal Paleozoic and Mesozoic strata cover the basement complex except in the Grand Canyon area, and (2) the vast platform made up of schist and granite which composes the southwestern half of the State (Fig. 1). The trend of the main Paleozoic waterways on this platform was toward central Arizona from the northwest and, approximately along the mountain belt of Ransome, from the southeast. Throughout the southwestern region stratigraphic contacts and interrelation of rocks can be studied with minutest detail, whereas the paleogeography of northeastern Arizona, with much evidence buried beneath the plateau, can only be surmised from the

studies of the rocks along its southwestern rim and in the Grand Canyon.

L. F. Noble (1914) studied the Archeozoic basement complex and Algonkian strata of the Grand Canyon area and, in the words of Ransome (Noble, 1914, p. 8),

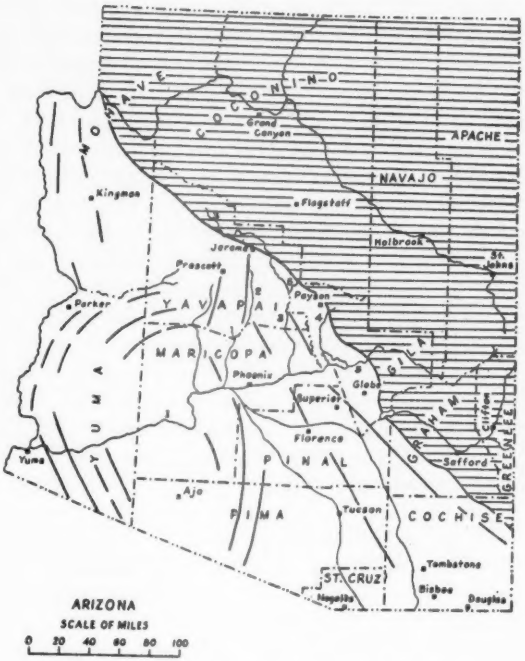

FIGURE 1.—*Orientation map of Arizona*

Basement platform unshaded, trend and virgation of the schistose and granitic mountains indicated by black lines. Plateau region with nearly horizontal strata ruled. (1) Gila River, (2) Agua Fria, (3) Verde River, (4) Tonto Creek, (5) Salt River, (6) Natural Bridge.

"has not only thrown light on the pre-Cambrian history of the Colorado Plateau region, but has supplied geologists who are working in the southwestern part of the country with a standard of comparison for Algonkian strata exposed elsewhere in that region."

The major facts that seem to be firmly established in the pre-Cambrian geology of the Grand Canyon are that (1) only the Archean rocks were affected by the great granitic intrusion, (2) a profound erosion interval separated the Archeozoic rocks from the Algonkian, and (3) a profound

erosion interval with structural adjustments elapsed before the introduction of Cambrian waters.

Two contributions have been made recently regarding pre-Cambrian rocks and geological events of central Arizona, paleogeographically the most important part of the State.

N. E. A. Hinds (1935, p. 27-29; 1936, p. 92-102; 1937, p. 1994-1998) outlined the geological development in central Arizona during the pre-Cambrian which he based partly on previous, and partly on his own, observations. As in the Grand Canyon, the Archeozoic rocks had been tilted, intruded by granite, and eroded before Algonkian time. In Algonkian time eruptions of rhyolite preceded the deposition of the Mazatzal quartzite, one of the principal pre-Cambrian members of central Arizona; orogeny, another granitic intrusion, and erosion followed.

Wilson (1939), who studied the central Arizona area in detail, offered a different interpretation. Volcanics, Yaeger greenstone, and Red Rock rhyolite are at the base of the pre-Cambrian complex. On the rhyolite are partly metamorphosed shale and grit of the Alder series which Wilson (1939, p. 1122) correlated with the sedimentary portion of the Yavapai schist of central Arizona. The Yavapai schist is usually placed with the Vishnu schist of the Grand Canyon and the Pinal schist of southeastern Arizona in the pre-Cambrian (Jaggar and Palache, 1905; *compare* Wilmarth, 1938, p. 2383), older pre-Cambrian (Ransome, 1916), or pre-Cambrian crystalline rocks (Darton, 1925). Above the Alder series more quartzite and shale is followed by the Mazatzal quartzite which, however, in the headwaters of the East Verde River and of the Tonto Creek rests directly on the Red Rock rhyolite. Wilson found evidence for only one granitic intrusion in central Arizona—of post-Mazatzal quartzite time; he gathered ample proof of the effect of a granitic intrusion on both the Red Rock rhyolite and the Mazatzal quartzite.

Wilson's evidence is opposed to the development of the pre-Cambrian complex in other regions of Arizona. The difficulty in his interpretation is that, like the Laurentian granite, the Yaeger greenstone and Red Rock rhyolite do not rest on any known foundation, but unlike that classic granite they are so limited areally as to cause doubt about their role, which they may be construed to share with the Archeozoic schists, of the most primordial rocks of Arizona; they have no counterparts in the Grand Canyon or elsewhere in the State.

The studies of geological sections in various parts of the Grand Canyon have proved the existence, within Arizona, of pre-Algonkian granite. The Algonkian rocks there, between the Archeozoic schist and Cambrian sedimentaries, have not been affected by the granite which intruded the schist.

Wilson's data accumulated in central Arizona admit of three theoretically possible solutions: (1) The Mazatzal quartzite, together with the older greenstones, rhyolites, and the equivalent of the Yavapai schist, all intruded by the same granite, is older than the Archeozoic Vishnu and Pinal schists and is one of the oldest members of the basement

TABLE 1.—*Pre-Cambrian of Arizona. Interpretation* 1

Southeastern Arizona		Central Arizona			Northern Arizona
Cambrian	Cambrian	Cambrian	Devonian		Cambrian
Unconformity					Unconformity
Apache group				Apache group	Grand Canyon series
Ep-Archean granitic intrusion and erosion interval					
Pinal schist					Vishnu schist
		Mazatzal group			

complex of Arizona, in which case Wilson's contention of only one major pre-Cambrian granitic intrusion is sustained (Table 1). Against the Archeozoic age of the Mazatzal quartzite, however, is the fact that in many places the Mazatzal is in no way more altered than the similar sediments of the Grand Canyon series, for instance the normal conglomerates within it in the East Verde headwaters. (2) The Mazatzal quartzite is younger than the Archeozoic schist but older than the strata of the Grand Canyon series. Assuming again only one major granitic intrusion it follows that the development of the Mazatzal quartzite with preceding volcanic activities is a chapter of the pre-Cambrian history not represented in the Grand Canyon area or elsewhere in Arizona. Such a postulation would call for a widespread removal of these or equivalent rocks, all intruded by granite simultaneously with the Archeozoic sedimentaries, with the evidence of their existence preserved only in a limited area of central Arizona. This alternative, however, admits either of one (post-Mazatzal) or, as proposed by Hinds, of two (post-Vishnu and post-Mazatzal) granitic intrusions and erosion intervals (Table 2). (3) The Mazatzal rocks—*i.e.*, volcanics, schist, quartzite, and intrusive granite—

are of the same age as, or younger than, the Grand Canyon series but represent an extremely localized development confined to central Arizona (Table 3). The Mazatzal quartzite overlaps the older sedimentary rocks and in the Payson area of central Arizona rests with a basal conglomerate directly on the Red Rock rhyolite (Wilson, 1939, p. 1125);

TABLE 2.—*Pre-Cambrian of Arizona. Interpretation 2*

Southeastern Arizona		Central Arizona			Northern Arizona
Cambrian	Cambrian	Cambrian	Devonian		Cambrian
Unconformity					Unconformity
Apache group				Apache group	Grand Canyon series
Post-Mazatzal granitic intrusion and erosion interval					
		Mazatzal group			
Post-Vishnu granitic intrusion and erosion interval ?					
Pinal schist					Vishnu schist

this may indicate a previous erosion with subsequent deposition of sediments and orogenic processes, culminating in a second granitic intrusion similar to other local and later granitic intrusions observed elsewhere in Arizona (Wilson, 1937, p. 1999).

The important fact in this connection is that, both in the Payson area of central Arizona at the headwaters of the East Verde River and at the Tonto Creek, and in the area north of Jerome in the basin of the main Verde River, the Cambrian Tapeats sandstone rests on the remarkably smooth surface of large masses of granite or on the uniformly tilted schist (Pl. 1, fig. 1). On the other hand, its contacts with the Mazatzal quartzite (Wilson, 1922, Pl. 31) are very much like the contacts of later Devonian strata with the same quartzite—the younger strata abut against or overlap the quartzite tilted at various angles within short distances. These two different relations of the Tapeats sandstone with the granite and the quartzite are strikingly similar to its relations with the smoothed surface of the Archeozoic and the rugged surface of the Algonkian of

the Grand Canyon area which have been recently described by Sharp (1940).

The writer is of opinion that the extensive granite, closely associated with schist, that forms the basement complex of the entire southwest of Arizona is pre-Algonkian.

TABLE 3.—*Pre-Cambrian of Arizona. Interpretation* 3

Southeastern Arizona		Central Arizona			Northern Arizona
Cambrian	Cambrian	Cambrian	Devonian		Cambrian
Unconformity					Unconformity
Apache group				Apache group	
		Post-Mazatzal granitic intrusion and erosion interval			
		Mazatzal group			Grand Canyon series
Ep-Archean granitic intrusion and erosion interval					
Pinal schist		Yavapai schist ?			Vishnu schist

Ransome (1916, p. 165-166) believed that a natural barrier, probably a submarine ridge, existed in central Arizona in Cambrian time and possibly throughout the Paleozoic; this concept has been gradually developed, with further research in stratigraphy and paleogeography, into the idea of a land mass—Mazatzal Land, a positive element that separated the two principal areas of deposition in Arizona all through the Paleozoic (Stoyanow, 1936, p. 462) and also had a definite bearing on the Mesozoic paleogeography of the State.

MARINE INVASIONS OF CAMBRIAN TIME

As early as late pre-Cambrian time the southeastern and northwestern areas of deposition seem to have been well established in Arizona. In Cambrian time, Mazatzal Land was encroached upon first from the north by the shallow Tapeats sea, the deposits of which are found farthest south of all three members of the Grand Canyon Cambrian, or the Tonto group, and are seen at present in the area of the East Verde headwaters

in central Arizona (Pl. 5a). The age of the Tapeats sandstone, formerly interpreted as very early Middle Cambrian (Stoyanow, 1936, p. 480), may be partly or entirely Early Cambrian, depending on the relation of the shales with *Olenellus* found in the western part of the Grand Canyon (McKee, 1937, p. 341) to the main body of the sandstone.

In *Anoria-Glossopleura-Alokistocare*, or earlier Middle Cambrian, time the northwestern sea retreated, and the corresponding formation, the Bright Angel shale, is not found very far south of the Grand Canyon. This retreat was followed by another transgression southward. The Muav limestone overlaps the Bright Angel shale, and at its southernmost observed termination, which is near Fort Rock, half way between the Grand Wash Cliffs and Prescott, it rests directly on the Tapeats sandstone.

The Cambrian waters began to invade southeastern Arizona only after these events. The Troy sea encroached from the southeast and reached the headwaters of the Tonto basin in central Arizona (Pl. 5b) where, at the present time, the Troy quartzite overlaps the rocks of the pre-Cambrian Apache group and abuts against the Mazatzal quartzite. In southeastern Arizona the Troy quartzite likewise rests either on the Apache group or on the post-Apache effusives. The upper part of the Troy, often referred to as the Bolsa quartzite, overlaps the Troy quartzite proper on the pre-Algonkian platform composed of schist and granite, thus outlining the greatest extent of the Troy basin.

The northern boundaries of all consecutively younger Cambrian formations are regressive to the southeast, marking a gradual retreat of the Cambrian sea which was accompanied by the successive introduction of younger faunas. The late Middle Cambrian shaly Santa Catalina formation has not been observed north of the confluence of the Gila and San Pedro rivers (Pl. 5c), and the somewhat younger Cochise formation with limestone beds which contain the *Olenoides pugio* [1] and *Eldoradia* faunas is even more restricted to the southeast and is not known far north of the Santa Catalina and Little Dragoon Mountains. Since these formations conformably overlie the Troy quartzite, the age of the latter is placed somewhere near *Elrathiella-Triplagnostus* time. (*Compare* Howell and Mason, 1938, p. 281.)

By the time of introduction of the *Tricrepicephalus texanus* and *Hesperaspis butleri* fauna, which characterizes the Upper Cambrian Abrigo formation, the sea had appreciably withdrawn farther southeastward, and the Abrigo formation is not found north of the Santa Catalina Mountains near Tucson (Pl. 5d). The strata with *Aphelaspis-Iddingsia*

[1] *Neolenus intermedius pugio* Walcott (1910, p. 35, Pl. 6, fig. 9). *See* Kobayashi (1935, p. 154).

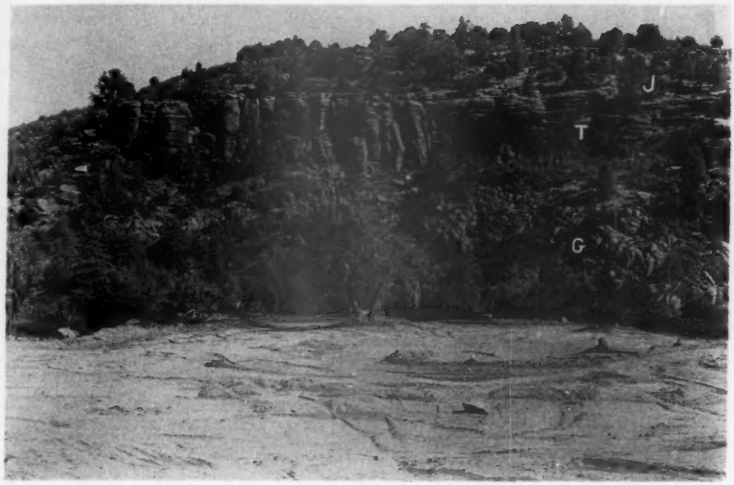

FIGURE 1. EAST VERDE CROSSING
Looking northwest. G, granite; T, Tapeats sandstone; J, Jerome formation with basal limestone resting directly on Cambrian.

FIGURE 2. TONTO CREEK HEADWATERS
Looking north. D, base of Upper Devonian clastics with arthrodire fishes overlying R, Red Rock rhyolite.

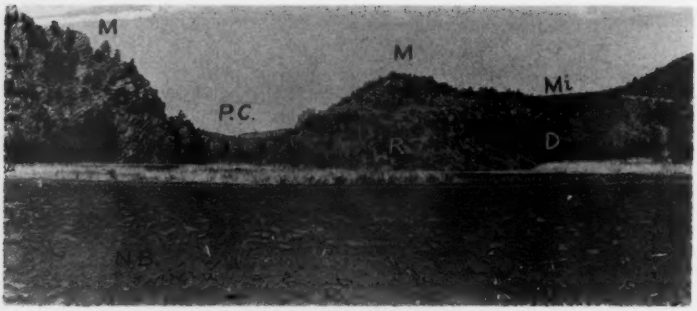

FIGURE 1. NATURAL BRIDGE
Looking north. Pine Creek, P.C., is deep in the gorge above and runs under the natural bridge, N.B., in foreground. R, Red Rock rhyolite; M, Mazatzal quartzite; D, basal Devonian sandstone; Mi, Mississippian Redwall limestone.

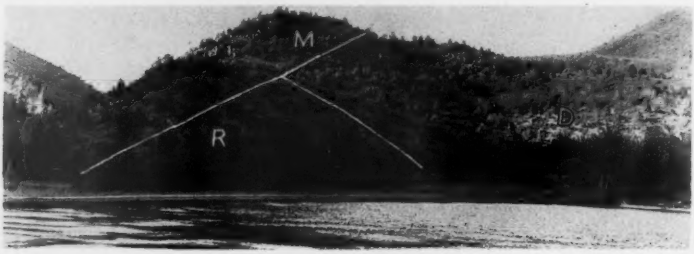

FIGURE 2. DETAIL OF FIGURE 1
East of Pine Creek. R, Red Rock rhyolite; M, Mazatzal quartzite; D, basal Devonian sandstone abutting against Algonkian rocks.

FIGURE 3. NATURAL BRIDGE
East side of Pine Creek gorge below settlement. D, basal Devonian sandstone on R, Red Rock rhyolite.

PALEOZOIC — PRE-CAMBRIAN CONTACTS AT NATURAL BRIDGE, CENTRAL ARIZONA

assemblages (Rincon limestone and Peppersauce Canyon sandstone) have approximately the same areal extent, whereas the beds with *Chariocephalus, Idahoia, Irvingella,* and *Maladia* (Copper Queen limestone) are restricted to the very southeastern corner of the State.

It is concluded from the foregoing discussion that in Cambrian time there was no communication between the northwestern and southeastern depositional areas of Arizona. The absence of the late Middle Cambrian and Upper Cambrian in northern Arizona, as seen from the sections in the Grand Canyon, rather supports this contention. Recently E. D. McKee located strata with a *Crepicephalus* fauna northwest of the Grand Canyon area in the Muddy Mountains, Nevada, and in the Virgin Mountains which are partly in northwestern Arizona (personal communication, September 12, 1941). The northwestern basin was in open communication with the Cordilleran geosyncline. On the other hand, the Cambrian southeastern basin was not directly connected with the Cordilleran sea across Arizona and southeastern California, but there is ample evidence that the Paleozoic seas of southeastern Arizona trough deepened southeastward into Mexico.

The presence of *Olenoides* and *Eldoradia* faunas in the late Middle Cambrian of southeastern Arizona needs an explanation since the nearest known locality with *Olenoides* is the House Range, Utah; the nearest localities with *Eldoradia* are the Eureka district, Nevada, and the Marble Mountains, California. Theoretically the intercommunication was possible either north of Arizona through the present Colorado Plateau via a Ouachita waterway east of Mazatzal Land or south of it through Sonora. Against the first possibility is the absence of Middle Cambrian strata in the Texan part of the Ouachita trough and the lack of evidence of the *Olenoides-Eldoradia* faunas in the Cambrian of New Mexico.

In his earlier interpretation the writer was inclined to regard the Middle Cambrian strata of the Marble Mountains near Cadiz, San Bernardino County, California (Hazzard and Crickmay, 1933, Map 1), as the southernmost termination of the Cordilleran sea and postulated a view (Stoyanow, 1936, p. 462) that Mazatzal Land might have extended from southwestern Arizona across southeastern California into the present area of Peninsular ranges. Important data on the stratigraphy of Sonora have been obtained recently by two able Mexican geologists, Isauro G. Gomez L. and L. Torres. The collected paleontological material has been brought to the University of Arizona for identification. Messrs. Gomez and Torres very kindly provided the writer with valuable field observations. In the Rio Magdalena Valley, northwestern Sonora, early Middle Cambrian strata with *Glossopleura, Alokistocare,* and *Anoria?* are present. This is the Arrojos formation, an equivalent of the Bright Angel

shale of the Grand Canyon. The type locality, Arrojos Hills, south of Caborca, is 50 miles east of the Gulf of California and 400 miles southeast of Cadiz, California, where similar faunas are known (Darton, 1907, p. 470; Clark, 1921, p. 6; Resser, 1928, p. 10; Hazzard and Crickmay, 1933, p. 74; Mason, 1935, p. 97; Hazzard and Mason, 1936, p. 233). The entire Paleozoic trough of northwestern Sonora, as inferred from the distribution of studied outcrops, extends southeastward for an unknown distance and is separated from the trough of southeastern Arizona and northeastern Sonora by a wide belt of Archeozoic schist and granite which connects to the northwest with the similar rocks of southwestern Arizona. This is the Altar Headland of Mazatzal Land. The Arrojos formation consists of about 900 feet of alternating shale and limestone. It rests unconformably on the Jojoba formation which is made up of massive limestone beds with reefs of *Collenia* and is provisionally placed in the Algonkian. The Arrojos formation appears to be overlain by Mississippian limestone.

The significance of this paleontological find is that it establishes an early Middle Cambrian invasion in Sonora from the Cordilleran Sea or the Pacific Ocean as shown on an early map of Schuchert (1910, Pl. 52). The northwestern trend of this seaway was, however, to the west of that postulated by Schuchert if the localities with the *Glossopleura-Alokistocare* fauna in California, Nevada, and Arizona are considered. It is indicated as extending from Caborca, Sonora, through the Marble, Providence, and Nopah-Resting Springs Mountains in California, the Goodsprings quadrangle and the Muddy Mountains in Nevada, and northward along the Cordilleran geosyncline with an embayment or cul-de-sac into northern Arizona.

It is plausible to suppose that late Middle Cambrian waters entered southeastern Arizona through Sonora south of the Altar Headland. The *Glossopleura-Alokistocare* zone is about 200 feet above the base of the Arrojos formation, and younger faunas may be present within the latter. However, there is no evidence of the presence of the late Middle Cambrian *Olenoides* fauna like that of southeastern Arizona in any of the Cambrian localities of southeastern California.

Probably even in Late Cambrian time the southeastern Arizona basin did not belong entirely in the Ouachita trough (*compare* Schuchert and Dunbar, 1941, p. 126, Fig. 70) and a communication existed between the Cordilleran and Ouachita seaways through Sonora (*compare* Schuchert, 1923, p. 217, Fig. 5) with an arm into southeastern Arizona. The Upper Cambrian trilobite fauna of southeastern Arizona associated with and following *Tricrepicephalus texanus* has but a few common genera with that of Texas. *Chariocephalus, Hesperaspis,* and *Maladia* thus far

are known only from Arizona, whereas *Burnetia, Conaspis, Pterocephalia,* and *Wilbernia* are found only in Texas; *Aphelaspis, Iddingsia, Irvingella, Tricrepicephalus,* and probably *Idahoia* (Bridge, 1937, p. 256) are common to the Upper Cambrian of both States.

DEVELOPMENT OF MAZATZAL LAND IN EARLY PALEOZOIC TIME AND LATE DEVONIAN TRANSGRESSION

The northwestern shore line of the southeastern Cambrian basin of Arizona lay progressively farther to the southeast as the sea retreated from Mazatzal Land. However, more of that land rose above the sea through the Ordovician, Silurian, and earlier Devonian. The only described Ordovician strata of Arizona, the Longfellow limestone of Beekmantown age (Lindgren, 1905, p. 65; Stoyanow, 1936, p. 483) in the Clifton-Morenci mining district and the unnamed limestone of the same age (Gilbert, 1875, p. 512; C. A. White, 1875, p. 77; Darton, 1925, p. 53; Stoyanow, 1936, p. 481) in the Dos Cabezas Mountains are both in southeastern Arizona near the New Mexico border (Pl. 5e). The exact age of the Coronado quartzite of the Clifton-Morenci area (Lindgren, 1905) is still unknown. It is reasonable to surmise, however, that the deposition of this quartzite did not correspond with the regressive phase of the Late Cambrian sea in southeastern Arizona. It probably marks the initial stage of the brief Early Ordovician transgression which eventually culminated in the formation of the fossiliferous Longfellow limestone.

West of this limited depositional area of Early Ordovician time there are neither Ordovician nor Silurian strata within Arizona, and none are definitely established in the Muddy Mountains, Nevada (Longwell, 1928, p. 22), in the Goodsprings quadrangle, Nevada (Hewett, 1931, p. 13), and south of Beaver County in Utah (B. S. Butler, 1920, p. 80, Pl. 4).

The general southward and eastward thinning of the Ordovician and Silurian formations in eastern Nevada and western Utah has been summarized in correlation tables prepared by Nolan (1935, p. 24). In an earlier paper Nolan (1928, p. 154) stressed the fact that the early Paleozoic sediments thicken westward toward the axis of the Cordilleran geosyncline from southwestern Utah and from Arizona—that is, from Mazatzal Land; on the attached map he marked the eastern limit of thick early Paleozoic sedimentation of the Cordilleran geosyncline and the western boundary of the pre-Carboniferous Paleozoic positive element, in this way outlining Mazatzal Land on the west.

In southeastern California Ordovician strata are present in the Nopah-Resting Springs Mountains (Hazzard, 1937, p. 322). The areas with Upper Ordovician limestone beds near La Casita in central Sonora (Dum-

ble, 1900, p. 148; Schuchert, 1935, p. 135) are, according to R. E. King (1939, p. 1641, Pl. 1), between 28.30 N. Lat. and 29.00 N. Lat., east and west of 110.00 W. Long., about 150 miles south of the international border.

All available evidence indicates that the Ordovician and Silurian seas were in contact only with the extreme marginal regions of Mazatzal Land.

The Early and Middle Devonian waters bounded Mazatzal Land on the north and northwest considerably beyond the borders of Arizona, as is inferred from C. W. Merriam's (1940, p. 4) index map of better-known Devonian sections of western North America from the Inyo Mountains, California, northeasterly to Randolph, Utah. The presence of Middle Devonian fauna in the Goodsprings quadrangle, southeastern Nevada (Hewett, 1931, p. 15), and in the Nopah-Resting Mountains, southeastern California (Hazzard, 1937, p. 330), has not been definitely established. No Devonian strata have been reported from Sonora west of the Altar Headland either by previous or recent explorers.

The Late Devonian waters entered Arizona from the northwest and southeast, following in a general way the trend of Cambrian ingressions (Pl. 5f). Both in the extreme northwest and southeast of Arizona the basal Upper Devonian limestone rests directly on the underlying Cambrian strata with apparent conformity. This is not so in central Arizona where, especially in the Payson area between the headwaters of the East Verde River and the Tonto Creek, the Late Devonian waters, spreading from the northwest and the southeast, encountered a greatly eroded land made up of the pre-Cambrian schist, volcanics, Mazatzal quartzite, and granite. The granite, capped by the horizontal Cambrian Tapeats sandstone, occupied broad depressions, whereas the highlands were composed of the tilted Mazatzal quartzite and Red Rock rhyolite. Accordingly, the Devonian sea encroaching from the northwest deposited thin-bedded limestone directly and with apparent conformity on the Cambrian sandstone in the lowlands, but the elevated areas were vigorously attacked by the surf with the formation of basal conglomerates and coarse sandstones which contain remains of arthrodiran fishes. Later on, with further erosion and submergence, the clastic material was also carried into the depressions, and the basal limestone there was overlain by finer sandstone and shaly limestone, also with fish remains (Fig. 2). After this, with frequent but brief oscillations, the rest of the Jerome formation, culminating in coral reefs with brachiopod fauna, which extends from the East Verde headwaters northwestward beyond Jerome, was deposited (Stoyanow, 1936, p. 495).

Many incorrect interpretations have been made regarding the geology

of this area, and deciphering of its paleogeography and stratigraphy is here briefly summarized.

Ransome (1916, p. 159) described a sequence of Paleozoic strata between Payson and Pine near Natural Bridge, Arizona, on Pine Creek, a

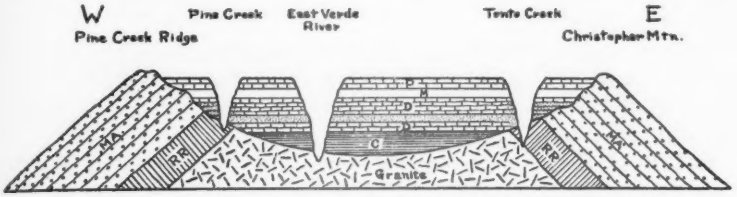

FIGURE 2.—*Sketch section across Payson Headland from Pine Creek Ridge to Christopher Mountain*

RR, Red Rock rhyolite; MA, Mazatzal quartzite; C, Cambrian; D, Devonian; M, Mississippian; P, Pennsylvanian. Faulting omitted. Distance 25 miles.

confluent of the East Verde River. The nearly horizonal limestone beds, resting, as Ransome thought, on the Cambrian sandstone (Tapeats?), abut against the tilted Mazatzal quartzite underlain by a crystalline rock which Ransome took for granite but which Wilson later showed to be a volcanic rock, the Red Rock rhyolite. Ransome did not differentiate the limestone deposits which rest on the supposed Tapeats sandstone. He was impressed by the proximity of this series of nearly horizontal Paleozoic strata to the upturned pre-Cambrian Mazatzal quartzite and also by the fact that certain Paleozoic members contained angular fragments of this quartzite. He thus concluded that:

"While it is difficult to understand the survival of a small sharp ridge as an isolated monadnock on a surface so nearly worn down to a plain as that over which the arenaceous deposits of the Cambrian sea were spread for thousands of square miles, it is still more difficult to account in any other way for this island-like mass of indurated tilted strata surrounded by dissimilar, nearly horizontal, and apparently younger beds."

Later E. D. Wilson (1922, p. 305) noticed that the basal Paleozoic sandstone in the Natural Bridge region differed lithologically from the Tapeats:

"Due to the nature of the old uneven surface, the Tapeats sandstone is for the most part absent . . . however, there is usually between the pre-Cambrian rocks and the true 'Redwall' limestone a member which in different localities varies between limestone, sandstone, and arkose, that might sometimes be mistaken for the Tapeats formation. Whether this always represents merely the base of the 'Redwall' limestone, or sometimes the Tapeats sandstone as well, is difficult to say."

In the geological map of Arizona, published in 1923 by the Arizona Bureau of Mines in cooperation with the U. S. Geological Survey, the basal Paleozoic strata of Natural Bridge were mapped as Cambrian.

Darton (1925, p. 44, 249) followed Ransome in interpreting the same rock of Natural Bridge as the Tapeats sandstone.

In 1925 Schuchert and the writer found arthrodiran fishes between Payson and Pine, in a sandstone lithologically very similar to the Tapeats; they concluded that the entire sandstone might be Devonian (Stoyanow, 1926, p. 312).

Ten years later Hinds (1935, p. 28) referred to the basal conglomerate and sandstone of the Natural Bridge area as Tapeats and described some effects of erosion of the surface of Mazatzal Land that, in his opinion, had taken place in the pre-Tapeats time.

In 1936 the writer criticized Ransome's theory of monadnock in the upper part of the Pine Creek Canyon and explained the abutting of the horizontal Paleozoic limestones, which he had separated earlier (Stoyanow, 1926, p. 314) on faunal evidence into Devonian, Mississippian, and Pennsylvanian, by a fault (Stoyanow, 1936, p. 516-517). He discussed the fish-bearing deposits of the Payson-Pine area and their lithological relation to the Tapeats sandstone but placed the basal Paleozoic beds of Natural Bridge in the Cambrian in accordance with the views of Ransome, Darton, and Hinds.

Only in recent years has the interrelation of the strata in the headwaters of the East Verde River and the Tonto Creek come to light. The basal platform of the Payson area, referred to in this paper as the Payson Headland, is made up of schist and granite. On the north, the granite of this headland is overlain by almost horizontal Tapeats sandstone on which conformably rest the thin-bedded limestones, the basal members of the Upper Devonian Jerome formation (Stoyanow, 1936, p. 495-500) (Pl. 1, fig. 1). The outcrops of the Mazatzal quartzite at Natural Bridge are at the western margin of the headland, while on its eastern margin, in the headwaters of the Tonto Creek, 25 miles away, are cliffs also made up of the greatly disturbed and tilted Mazatzal quartzite like those of Natural Bridge (Fig. 2).

In the Tonto Creek headwaters, the Mazatzal quartzite, Red Rock rhyolite, and in places also the granite are overlain by the same basal conglomerates and sandstones, composed partly of chunks and boulders of the underlying rocks, as are present at the base of the Paleozoic sequence at Natural Bridge. In these basal beds the writer found, in the summer of 1939, abundant remains of arthrodiran fishes (Pl. 1, fig. 2).

Another reconnaissance in the Natural Bridge area revealed that independently of whether the contact between the Mazatzal quartzite and the abutting Devonian limestone on the east side of Pine Creek—the original location of Ransome's studies discussed above—might have been

caused by local faulting (Pl. 2, figs. 1, 2), on the west side of the creek the basal conglomerates and sandstones, conformably underlying the younger Devonian strata, unmistakably rest upon the tilted Mazatzal quartzite (Fig. 3; Pl. 3, figs. 1, 2). On the east side of the creek, but

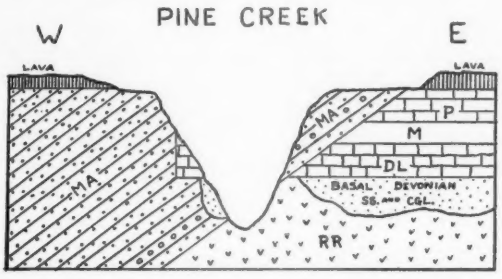

FIGURE 3.—*Sketch section across Pine Creek above the Natural Bridge settlement*

RR, Red Rock rhyolite; MA, Mazatzal quartzite; DL, Devonian limestone; M, Mississippian; P, Pennsylvanian.

below the Natural Bridge settlement, the same basal conglomerates rest on the rhyolite (Pl. 2, fig. 3).

These studies prove that the coarse basal conglomerates with angular fragments of the Mazatzal quartzite, including the large boulders of talus and fan debris described by Hinds (1935, p. 28) and others on Pine Creek (Pl. 3, fig. 3), are Upper Devonian and not Cambrian, and the accompanying erosion on Mazatzal Land, therefore, was coincident with the Late Devonian and not with the Cambrian transgression.

The basal Devonian clastics, where they rest on the Mazatzal quartzite, differ markedly from the Tapeats sandstone which is slabby whether it overlies the granite, as in the Payson Headland, or rests on the greenstone and schist, as it does farther northwest in the Jerome-Del Rio area.

As stated above, the basal complex of the central part of the Payson Headland is schist and granite. On the west, southwest, and east are isolated masses of the Mazatzal quartzite. The eastern margin of the headland most effectively separates the rocks of the southeastern area of deposition from those of the northwestern basin. This is the location of Ransome's barrier or "submerged ridge" which prevented the northwestward advance of the Apache and Troy waters. On the east, east of the marginal Christopher Mountain massive made up of the Mazatzal quartzite, in the deep canyons cut by Gordon and Haigler creeks, tributaries of Tonto Creek, the unconformable contact between the Mazatzal quartzite and the beds of the Apache group and the Troy quartzite is

seen. These contacts were originally described by Ransome (1915, p. 116; 1916, p. 157) and later by E. D. Wilson (1922, p. 307; 1939, p. 1148) and by Darton (1925, p. 234).

The pre-Mazatzal and Mazatzal rocks of central Arizona, which in late Algonkian time stopped the Apache waters on the northwest, limited the Tapeats sea on the southeast and the Troy basin on the northwest in Cambrian time, formed, together with the overlying Cambrian strata, a headland which was attacked and encroached upon by the seas both from the northwest and from the southeast in Devonian time.

The exact nature of this Devonian headland in the northeastern direction, beneath the Colorado Plateau, is, of course, unknown, but, since the lithology, stratigraphic sequence, and marine faunas of the Jerome formation northwest of it are essentially different from those of the Martin formation to the southeast, communication between the two depositional areas was probably considerably impeded. Along the marginal, predominantly arenaceous, facies, however, the arthrodiran fishes existed from Superior to Jerome, and from there, in a more calcareous facies, to Flagstaff (Pl. 5f).

The Devonian on Tonto Creek is almost entirely restricted to the fish-bearing clastics which rest on the Mazatzal quartzite and are overlain directly by the Mississippian Redwall limestone. The *Cladopora* reefs with brachiopod fauna of the Jerome formation, which extend from Payson to Jerome and some distance north of it, are not developed on Tonto Creek.

In concluding this brief outline of the shaping of the Payson Headland and the relative role played in it by the basal Cambrian and Devonian strata, the writer wishes to comment on a tendency in recent literature to discredit in their original meaning the terms "Tonto group" and "Tapeats sandstone," which are extensively used in the Cambrian stratigraphy of northern Arizona. In an article entitled *Tonto group of Arizona is what? and where?* C. R. Keyes (1940, p. 237) placed the Tonto Creek near Fort Apache, in Navajo County, over 75 miles away from Gilbert's (1875, p. 60) type locality in the true Tonto basin, which is in the western part of Gila County, and claimed that the Apache group of Ransome is the original Tonto group. Keyes' assumption is based on Marvine's erroneous reference to the quartzites of the Apache group and the Troy quartzite in the upper basin of the Salt River as the "Tonto sandstone" (Marvine, 1875, p. 219). It is true, although not mentioned by Keyes, that the Tapeats sandstone is not present anywhere in the Tonto Creek canyon proper, and if Gilbert or Marvine had in view the sandstone exposed on that creek it was the Devonian basal sandstone with arthrodiran fishes discussed above. However, since Gilbert origi-

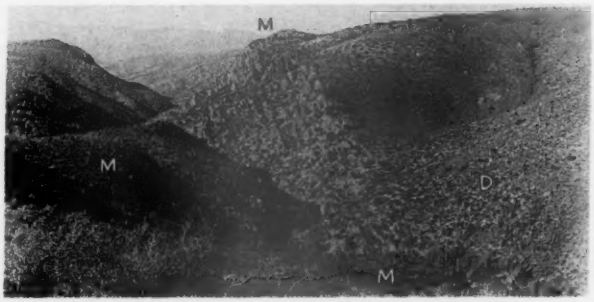

FIGURE 1. PINE CREEK GORGE ABOVE NATURAL BRIDGE
Looking south. D, Devonian sandstone resting unconformably upon M, Mazatzal quartzite, on the west side of Pine Creek.

FIGURE 2. WEST SIDE OF PINE CREEK GORGE ABOVE NATURAE BRIDGE
Looking south. D, basal Devonian sandstone unconformably overlying M, Mazatzal quartzite.

FIGURE 3. WEST SIDE OF PINE CREEK GORGE ABOVE NATURAL BRIDGE
Looking south. D, conglomeratic accumulations of basal Devonian material in pockets of M, Mazatzal quartzite.

CONTACTS BETWEEN DEVONIAN CLASTICS AND MAZATZAL QUARTZITE

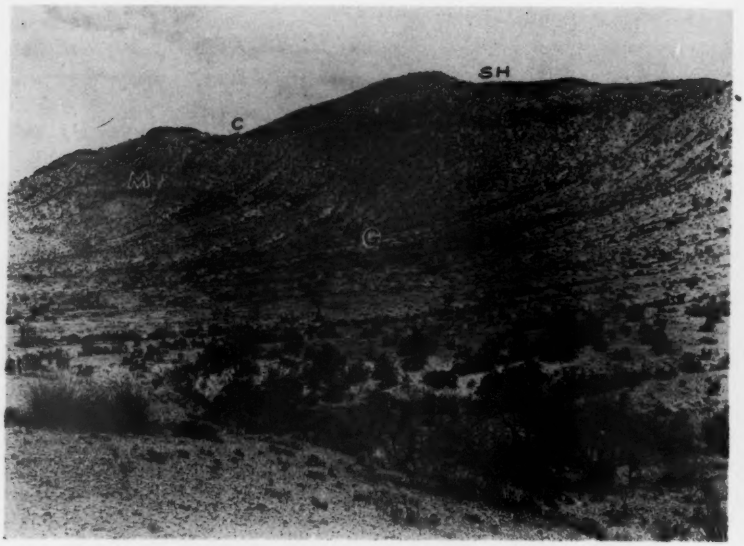

FIGURE 1. EAGLE BLUFF IN THE EMPIRE MOUNTAINS SOUTHEAST OF TUCSON
Looking northeast. G, gypsiferous series; M, basal limestone with Manzano fauna; C, contact between the Manzano beds and the Snyder Hill formation; SH, Snyder Hill formation.

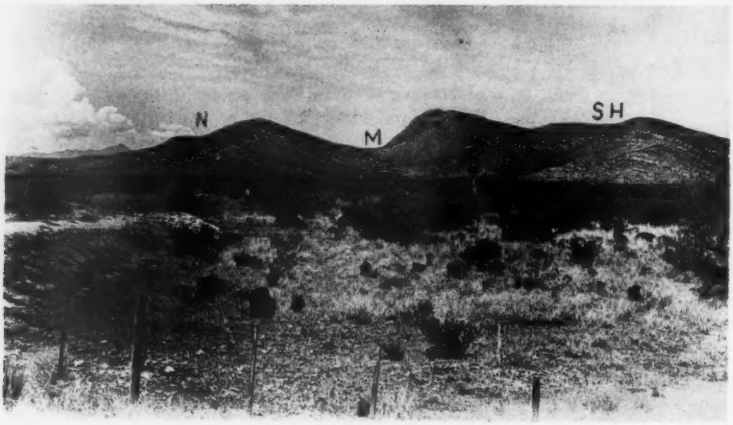

FIGURE 2. NACO HILLS WEST OF BISBEE
Looking west. N, Naco formation; M, Manzano beds, the gypsiferous series forms a saddle in the profile; SH, Snyder Hill formation.

nally referred to his "Tonto group" as located in the headwaters of the Verde and Tonto creeks, and because the western tributaries in the headwaters of Tonto Creek cut through Diamond Mesa and other mesas in which the Tapeats sandstone is well exposed, the presence of this sandstone, the basal member of the Tonto group, in the basin is beyond question. The fact that Marvine later confused the Apache beds with the Tonto group, or placed these beds within it, cannot invalidate the correctness of Gilbert's earlier definition.

In the northwestern basin the Devonian strata thin rather gradually in the northwestern direction from the Payson Headland as seen from the measurements at Jerome, Yavapai County (500 feet), and at Peach Springs, Mohave County (370 feet), but northeast of the axis of the basin these strata rapidly thin to 150 feet at Flagstaff, are not over 100 feet thick in the Grand Canyon area, and probably disappear northeast of the Grand Canyon.

The latest Upper Devonian strata, although different lithologically and faunally in the two depositional areas, indicate shallower waters. In the southeastern basin they are represented by the Lower Ouray formation, with *Camarotoechia endlichi*, which contains shale and sandstone in addition to limestone. This formation is known from the Little Dragoon Mountains near Benson to the upper basin of the Salt River, but it is absent in the Bisbee area in southeastern Arizona, where throughout the Paleozoic, but especially during the Late Devonian and Early Mississippian, the southeastern trough was deepest, and where the Devonian-Mississippian sequence consists, with slight local variations, only of limestone. In the northwestern basin, the Island Mesa beds likewise contain more shaly and arenaceous material than the underlying Jerome formation. These beds, with still undescribed fauna of heavy-shelled lamellibranchs and gastropods, are about 125 feet thick.

The small and isolated basin in the Clifton-Morenci area (Pl. 5f) is insufficiently known. Devonian age of its strata of argillaceous facies, the Morenci shale, was provisionally established by H. S. Williams on only one imperfectly preserved specimen of *Schizophoria* (Lindgren, 1905, p. 66-69). It is doubtful that this basin was directly connected with the southeastern trough.

The shallowing of the Devonian waterways toward the close of the period is clearly indicated all over Arizona except in the extreme northwestern and southeastern areas.

MISSISSIPPIAN SEAS AND PENNSYLVANIAN TRANSGRESSION

The Lower Mississippian strata of Arizona are well separated from the older rocks by a definite datum, the basal granular limestone which differs

markedly from the underlying Cambrian and Devonian sedimentaries. The Escabrosa limestone of the southeastern basin and the Redwall limestone of the northwestern depositional area are uniformly granular, crystalline, and predominantly composed of crinoidal material (Ransome, 1904, p. 42; Noble, 1922, p. 54; Stoyanow, 1936, p. 505).

The development of the Lower Mississippian waterway in southeastern Arizona, which coincided with the introduction of new faunas, was the sequel to a rather rapid deepening of the former Late Devonian trough and not to a new encroachment of the sea upon Mazatzal Land (Pl. 5g).

In the Bisbee area, the Escabrosa limestone rests conformably on the unfossiliferous (?) upper limestone beds of the Martin formation and farther northwestward on the Lower Ouray formation. Nowhere is there any evidence of interruption in deposition, no "profound erosional unconformity at its base, one of the great peneplanal surfaces of the continent" as Keyes (1940, p. 237) interprets it. The influence of Mazatzal Land on the southeastern sea in Early Mississippian time is seen in the gradual thinning of the Escabrosa limestone from Bisbee toward the Payson Headland and in the appearance, in the approaches to the headland, of the red regolithic material from the pre-Mississippian rocks of Mazatzal Land in the limestone beds. The thickness of the Escabrosa limestone north of Douglas is 950 feet (Stoyanow, 1936, p. 511), 800 feet at Bisbee (Ransome, 1904, p. 43), 425 feet at Superior (Harshman, 1939), not much over 150 feet at Lake Roosevelt (Stoyanow, 1936, p. 508), and much less than that in the areas closer to the Payson Headland (Darton, 1925, p. 237).

The western limit of the southeastern basin is not clear. Gilluly (1937, p. 40) described a fanglomerate in the Ajo area in which among other components limestone material, carrying Late Paleozoic fossils known in southeastern Arizona, was observed. W. L. Thomas reports a large number of middle-sized limestone boulders containing crinoids scattered on the floor of Yuma Desert southeast of Yuma. The redeposited material of Ajo undoubtedly was derived from the sedimentaries of the southeastern basin. In northwestern Sonora, west of the Altar Headland, the Devonian is unknown, and the Mississippian strata, which rest directly on the Cambrian, contain beds with Lithostrotiontidae, not represented in the Mississippian faunas of Arizona but recorded in the Cordilleran seaway in California. To the east, the Modoc limestone of the Clifton-Morenci area (Lindgren, 1905, p. 69; Stoyanow, 1936, p. 511) was deposited in a small, isolated water body apparently not connected directly with the larger southeastern basin.

In the northwestern depositional area, the Early Mississippian waters spread over the Devonian strata in a northeastern direction as is seen

from the overlap of the Redwall limestone on the Cambrian Muav limestone in the Grand Canyon where, according to McKee (1937, p. 341), the Redwall limestone thins eastward. The Redwall limestone in the Grand Canyon area is variously estimated as between 500 and 700 feet thick, depending on the interpretation of the age of the thinner limestone beds between the massive limestones of the Redwall and the base of the Supai formation. It is quite possible that, in the Grand Canyon or to the west of it, the Redwall limestone attains its maximum thickness within Arizona. Along the axis of the northwestern trough, the Redwall limestone thins to the southeast. In the Jerome area it is 300 feet thick, but it is only 80 feet thick farther southeast in the headwaters of the East Verde River (Stoyanow, 1936, p. 512, 513); in the upper basin of Tonto Creek the Redwall limestone has thinned to a few tens of feet. The earliest Lower Mississippian fauna of the southeastern depositional area is unknown in the Redwall limestone, from which it may be inferred that the Mississippian fauna, and probably the Mississippian waters, reached the northwestern basin somewhat later.

Toward the close of the Early Mississippian, the waters had completely withdrawn from the northwestern basin. In the southeastern basin, the withdrawal of the Mississippian sea was not so complete. In the southeasternmost part of Arizona the sea lingered through the Late Mississippian in a limited area in which the fossiliferous Paradise formation was laid down (Stoyanow, 1926; 1936; Hernon, 1935). This very limited area, in the Chiricahua Mountains north of Douglas, seems to have been the only part of Arizona under the sea in Late Mississippian time (Pl. 5h).

The marine cycle under discussion began with the Late Devonian transgression from the northwest and southeast toward central Arizona; the sea was deepest in Early Mississippian time; the recession of the Early Mississippian waters, which followed almost precisely in the same manner and direction as did the withdrawing Cambrian seas, left the entire State bare, with the exception of the lingering Late Mississippian water body in its southeastern corner.

During the following Pennsylvanian transgression, the sea, issuing again from the extreme southeast of Arizona, reached the Payson Headland and encroached upon it, but the Pennsylvanian sea did not spread far beyond the headland northwestward and did not occupy the northwestern area of deposition, the site of the previous seas in northwestern Arizona. Northwestern and northern Arizona was land through Pennsylvanian time with the possible exception of the extreme northwestern portion of the State in the neighborhood of the Grand Wash Cliffs and the Virgin River area (Longwell, 1921, p. 46; Reeside and Bassler,

1922, p. 66, 77; Longwell, 1928, p. 33; Longwell and Dunbar, 1936, p. 1203), since the evidence of Pennsylvanian faunas has been found in adjacent parts of Nevada and Utah. The Pennsylvanian water body of that region, however, belonged to a sea, the greater part of which was beyond the limits of Arizona and which at no time was in direct communication with the southeastern-central Pennsylvanian basin discussed in this paper (Pl. 5i).

In the extreme southeastern part of Arizona, the basal Pennsylvanian beds, like the basal beds of the Upper Devonian and the Lower Mississippian, are made up of limestone. They rest conformably on the Upper Mississippian Paradise formation in the Chiricahua Mountains, and with apparent conformity on the Lower Mississippian Escabrosa limestone at Bisbee. To the northwest, the first appearance of clastic rocks—intraformational conglomerates—within the basal limestone beds of the Naco, the Pennsylvanian formation of southeastern Arizona, has been recorded in the Empire Mountains about 45 miles northwest of Bisbee and 28 miles southeast of Tucson. At Superior, farther northwest, a well-defined conglomerate separates the Lower Mississippian strata from the basal Pennsylvanian limestones (Harshman, 1939). In the Payson area of central Arizona, the encroachment of the Pennsylvanian sea on Mazatzal Land is in full evidence. Basal and intraformational conglomerates contain boulders of the Mazatzal quartzite, thus indicating the progress of the invasion, and the distribution of Pennsylvanian strata shows that the sea had overcome the eastern barrier on the Payson Headland which was insurmountable for the Apache, Cambrian, Devonian, and Mississippian waters. The Pennsylvanian sea had spread over the Payson Headland to be checked only by the second line of the more resistant pre-Cambrian rocks at its western margin. The lithology of the Pennsylvanian strata on the headland is strikingly different from that in the areas southeast of it. Clastic rocks predominate. The thin, slabby, fossiliferous limestone beds in the lower part of the Pennsylvanian sequence are underlain and overlain by maroon shale, and in its upper part are brown sandstones and grayish thin-bedded and laminated shales of considerable thickness, as seen from the outcrops in the Tonto Creek headwaters. Such rocks are unknown in the Naco and Galiuro, the Pennsylvanian of southeastern and central Arizona.

LAND AND SEA IN PERMIAN TIME

The writer has shown that the southeastern basin was deepest in its extreme southeastern end through each of the outlined Paleozoic invasions upon Mazatzal Land. Within southeastern Arizona, every transgression started from the southeast, and during the withdrawals the

waters either stayed there longer or lingered until the time for a new northwestward advance arrived. This was the area of the greatest accumulation of Paleozoic limestone deposits in the State.

The northwestern trough was permanently destroyed with the close of the Early Mississippian, and no broad marine bodies occupied northern Arizona through the Pennsylvanian and the greater part of Permian time until the development of the Kaibab facies there at the termination of the Permian submergence in Arizona.

In southeastern Arizona, however, deeper-sea deposition went on from the Pennsylvanian into the Permian with only brief interruptions.

In pre-Kaibab time, and especially during the deposition of the earlier Supai, the broad, shallow brackish and fresh-water bodies of northern Arizona were controlled by the southeastern Permian sea, but the communication between the southeastern and northern regions was not through central Arizona. From time to time the marine waters of the southeastern basin reached, by brief incursions through eastern Arizona, east of Mazatzal Land, the shallow depressions where the deposition of the Supai red beds was going on (Pl. 5j).

The writer (1936, p. 521, 530) discussed the present status of the Upper Carboniferous stratigraphy in southeastern Arizona. When establishing the Naco limestone as a stratigraphic unit, Ransome (1904, p. 54) regarded it as prevalently Pennsylvanian. Girty (Ransome, 1904) tentatively separated the Naco into a lower part with well-known Pennsylvanian fauna and an upper part with a different assemblage of fossils predominantly composed of gastropods. It was pointed out by the writer that: (1) Within the total thickness of 3000 feet of Ransome's Naco, as seen in the Whetstone Mountains between Bisbee and Tucson, the Manzano fauna, as previously described by Girty (1909) from New Mexico and indicated for the upper part of the Naco by Darton (1926, p. 821), is present; (2) the gastropod fauna mentioned by Girty from Ransome's collections belongs in the Snyder Hill, a younger formation than the Manzano; and (3) the beds of Manzano age extend from southeastern into east-central Arizona, and limestone layers with Manzano fauna, the Fort Apache limestone, are found within the Supai formation near Fort Apache in the upper basin of the Salt River (Darton, 1925, p. 85; Stoyanow, 1936, p. 533). Consequently, the term "Naco" was restricted to the Pennsylvanian part of the original Naco limestone, and the presence of marine Permian strata in the southeastern basin was recognized.

In 1937 B. S. Butler and G. P. Sopp (1940) discovered a gypsiferous series in the Empire Mountains east of Tucson above the Pennsylvanian Naco limestone and below the Snyder Hill formation (Pl. 4, fig. 1). The

presence of gypsum below the Snyder Hill formation south of Tucson was mentioned before by C. F. Park (1929) and E. B. Eckel (1930). In the summer of 1940, the writer located the gypsiferous series and limestones with the Manzano fauna between the Naco and the Snyder Hill in the Naco Hills near Bisbee, the type locality of Ransome's Naco limestone. The Naco formation, rich in typical Pennsylvanian brachiopod and coral faunas, terminates in red shales and *Fusulina* limestones. The fusulinid specimens were submitted for examination to M. L. Thompson who kindly informed the writer (personal communication, March 6, 1941) that they are all referable to highly developed representatives of *Triticites* of upper Pennsylvanian rather than Permian aspect. They are comparable to the forms found in a zone at the top of the Magdalena in New Mexico. Above the Naco a sequence of shales, gypsum beds, and thin-bedded limestones aggregate 200 feet. These much softer beds form a saddle in the profile of the Naco Hills (Pl. 4, fig. 2). Within the upper part of these strata are limestone beds with the Manzano fauna. The term "Manzano" is applied here to the two discussed units in a broad sense, implying strata with the Permian Manzano fauna as described by Girty rather than any accepted or suggested divisions of the Manzano group in New Mexico. The Snyder Hill formation, one of the most resistant of the Arizona Paleozoic rocks and often the cap rock of mesas and buttes in southeastern Arizona, is at the top of Ransome's original "Naco limestone."

In a northwesterly direction, toward Tucson, the Permian basin was somewhat shallower in Manzano time; there are clastic rocks at the base of the gypsiferous series; the gypsum layers are more abundant and thicker; and the fossiliferous limestones in the upper part of the sections alternate with quartzites. These strata, between the Naco and the Snyder Hill, are now known in the Whetstone Mountains, Cochise County, and in the Empire Mountains, Pima County, southeast of Tucson; in the Santa Rita Mountains and in isolated hills between Tucson and Nogales, south of Tucson. In the Empire Mountains the gypsiferous series is 1460 feet thick, and the overlying fossiliferous limestones and quartzites are about 500 feet thick (measured by F. W. Galbraith). The gypsiferous beds in the Santa Rita Mountains are 1300 feet thick; the limestones with the Manzano fauna aggregate 900 feet (measured by W. R. Jones). The Manzano fauna is found invariably in the basal beds of the limestone series. It differs markedly from the faunal assemblages of the underlying Naco formation. The well-preserved silicified fossils of numerous, predominantly small shells of gastropods and lamellibranchs are easily leached out of limestone. Productids are exceedingly scarce in this facies. The most abundant diagnostic species are: *Orthonema*

socorroense Girty, *Phanerotrema manzanicum* Girty, *Nucula levatiformis* Walcott, *Astartella subquadrata* Girty, and *Schizodus wheeleri* Swallow? Girty, found with a number of closely allied forms. Farther north, in White River Indian Reservation, the Fort Apache limestone, with a similar Manzano fauna, rests on the Kinishba beds with a Permian flora, corroborating the Permian age of this molluscan fauna.

The Snyder Hill formation, with a novel gastropod fauna, apparently has a much more limited northward extension (Pl. 5k). With the gastropods occur dictyoclosti of the *ivesi-bassi* group, although not identical with the species from the Kaibab limestone, and also the characteristic forms of the *occidentalis* and *indicus* groups. The writer (1936, p. 532) separated under the name Chiricahua limestone higher beds in the extreme southeastern part of Arizona in which the Kaibab species, like *Aviculopecten coloradoensis* (Newberry) and *Dictyoclostus bassi* (McKee), are found. These strata, in which the presence of *Waagenoconcha montpelierensis* (Girty) has recently been established, are now known to extend west of Tucson. This unit is comparable to the Kaibab limestone, *sensu lato,* inclusive to the *Waagenoconcha montpelierensis* zone. Unlike the Kaibab limestone, however, the Snyder Hill formation and the Chiricahua limestone contain exotic species so far unknown in the Permian of the Southwest. Such, for instance, are a species of *Trachydomia* which is close to, though not identical with, *T. magna* Hayasaka (1938) from the Permian of Japan, and *Meekella* sp. of the group of *Meekella kueichowensis* Huang (1933, p. 31, Pl. 4, figs. 8-9) from the Permian of southwestern China.

The "Kaibab limestone" is heterogeneous and was deposited under oscillatory conditions which prevailed in northern Arizona during Permian time, as may be inferred from the rapid lateral and vertical variations of its members. Recently it has been divided by McKee (1938) into a lower unit, the Toroweap formation, and an upper unit, the Kaibab formation proper. This separation was established on the basis of an unconformity between the two formations and also on the vertical distribution of two species of dictyoclosti of the *Dictyoclostus ivesi-bassi* group. According to McKee's earlier interpretation (Stoyanow, 1936, p. 528, 529) *Dictyoclostus bassi* characterized the lower division, whereas *D. ivesi* was characteristic of the strata above the unconformity. With the progress of his research McKee firmly established the fact that the stratigraphic range of these two species is exactly reversed, *D. ivesi* being restricted to the Toroweap, and *D. bassi* to the Kaibab formation proper.

As has been mentioned above, the Snyder Hill formation is sharply distinguished from the underlying Manzano by the sudden advance of dictyoclosti of the *D. ivesi* and *D. occidentalis* types. In the polyphyletic

Semireticulati of the Arizona Permian *D. ivesi* and *D. bassi* are closely allied, as was recognized by McKee. Although there is no appreciable segregation of smaller forms referable to *D. ivesi* in the lower division of the Snyder Hill, the larger forms identifiable with *D. bassi* predominate in the upper Permian sequence of southeastern Arizona.

As pointed out by McKee, the Manzano molluscan fauna in northern Arizona is pre-eminently developed very high in the marine facies of the Permian strata. Thus, for instance, *Nucula levatiformis* and *Schizodus wheeleri* occur in the two upper divisions of the Kaibab—that is, in and above the *Waagenoconcha montpelierensis* zone (McKee, 1938, p. 161, 162, 244). In southeastern Arizona there is no marked penetration of the Manzano fauna into the basal Snyder Hill, and if there is a recurrence of such a faunal assemblage anywhere within the Snyder Hill formation and the Chiricahua limestone it has not yet been discovered.

Both the limited areal extent of the Snyder Hill and the Chiricahua limestone and the composition of their faunas suggest no direct communication between the southeastern basin and northern Arizona east of Mazatzal Land at the close of the Permian in the State. McKee (1938, p. 173) thinks that, during the development of the open-sea facies of the Kaibab, a communication might have existed between the depositional areas of northern and southern Arizona through western Arizona. There is no definite corroboration of this postulation at present, but there are certain facts that deserve consideration. Darton (1925, p. 215-223) described outcrops of limestone, closely associated with schist and granite, in the Harquahala, Plomosa, and Buckskin Mountains in western Arizona. He found some Carboniferous fossils in the less altered limestone beds of the Harquahala Mountains. He also noticed Carboniferous fossils in the redeposited boulders south of those mountains. In 1939 the writer examined post-Permian strata in the Bill Baer Hills south of the Harquahala Mountains. The strata which contain the boulders are arkosic sandstones with some quartz and alternate with thin-bedded quartzites. At many horizons of the sandstone are numerous, large and small, very angular boulders of blue limestone. The size and angularity of the boulders argue against extensive transportation. The fossils are satisfactorily preserved, and the fauna consists of crinoids, sea urchins, bellerophontids, and productids. This fauna is undoubtedly Permian but does not clarify its relation either to the Snyder Hill or the Kaibab faunas. On Plate 5k the Harquahala area is outlined with question mark as a possible site of a Permian basin of unknown relation. It is interesting, however, that in California, 1 mile north of the International Border and about the same distance west of the Colorado River, well-preserved silicified specimens of *Dictyoclostus occidentalis* and other

brachiopods of the Snyder Hill-Kaibab affinities are found in the lower beds of gravelly mesas along the All-American Canal survey line (R. Lenon, personal communication and specimens).

At present the stratigraphic position for an equivalent of the Coconino sandstone or of the Toroweap formation is not determinable within the continuous development of the open sea facies of southeastern Arizona represented by many hundreds of feet of an uninterrupted limestone sequence.

The relation of the Snyder Hill to the Kaibab formation also is not quite clear. McKee (1938, p. 35-54) has shown that each of the three members of the latter formation is composed of a number of localized facies. Facies 1 of the middle member rapidly wedges out south of Flagstaff. Facies 5 of the same member, with *Waagenoconcha montpelierensis*, is developed principally in Utah. In the upper member of the Kaibab the possible open-sea facies 4 is even more localized. For this reason no attempt is made here to date the Kaibab basin in relation to the seas of southeastern Arizona prior to the deposition of the Chiricahua limestone.

Plate 51 represents the rapidly decreasing Permian marine basins of Arizona toward the close of *W. montpelierensis* time. At that time the southeastern basin was in communication both with the Ouachita seaway and with the Pacific waters. I. G. Gomez and L. Torres (personal communication and specimens) found a Permian fauna with *W. montpelierensis* in Monos Hills (Monos formation), northwestern Sonora, about 50 miles east of the Gulf of California.

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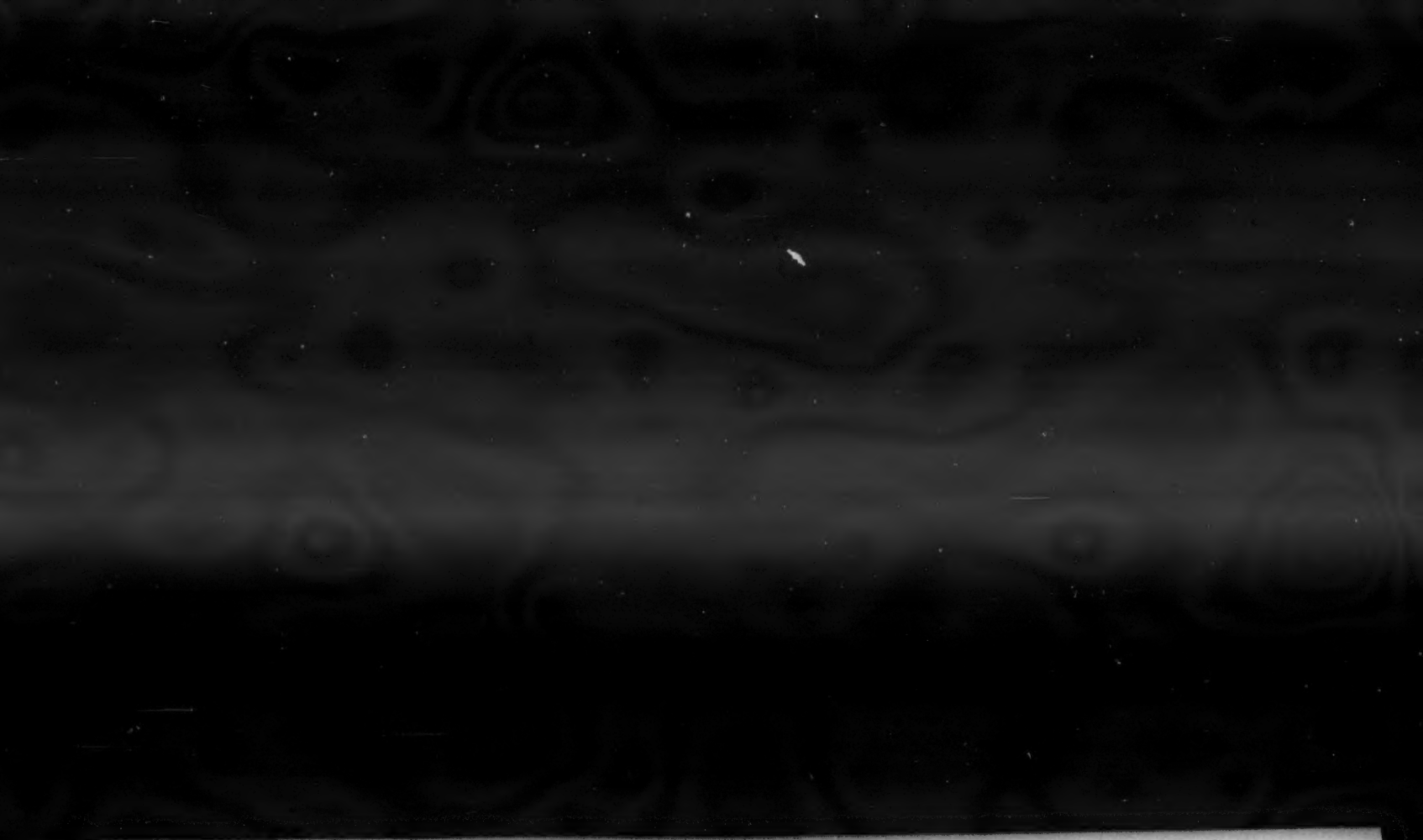
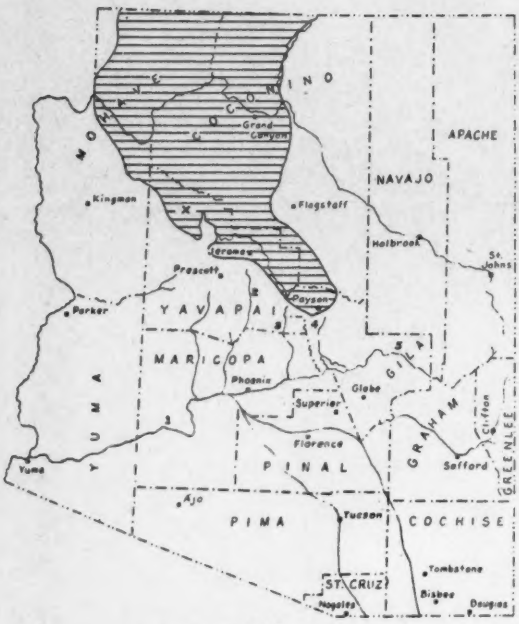
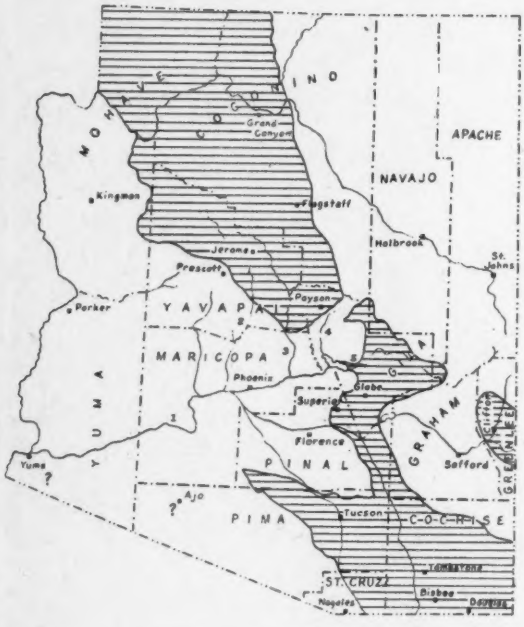




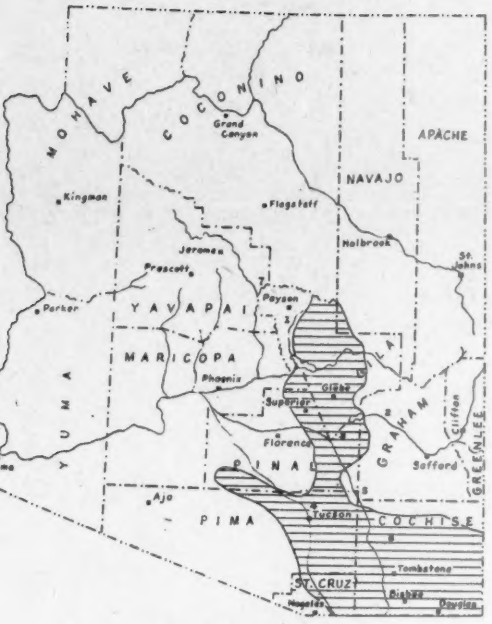
a.—*Early Middle Cambrian*

Tapeats basin ruled. Cross indicates southern limit of Muav basin. (1) Gila River, (2) Agua Fria, (3) Verde River, (4) Tonto Creek, (5) Salt River.



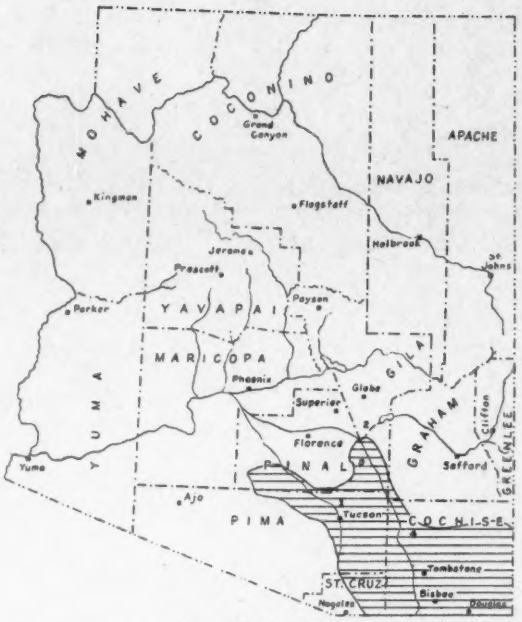
g.—*Early Mississippian*

Redwall, Escabrosa, and Modoc basins ruled. (1) Gila River, (2) Agua Fria, (3) Verde River, (4) Tonto Creek, (5) Salt River.



b.—*Late Middle Cambrian*

Troy basin ruled. (1) Tonto Creek, (2) Gila River, (3) Galiuro Mountains, (4) Santa Catalina Mountains, (5) San Pedro River, (6) Little Dragoon Mountains, (7) East Verde River.



c.—*Late Middle Cambrian*

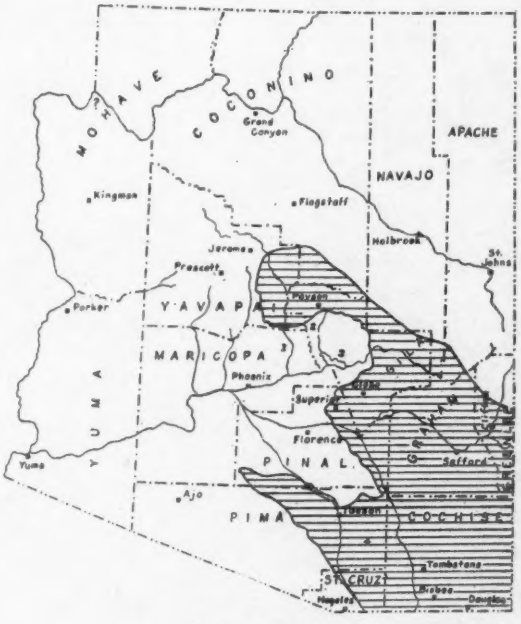
Santa Catalina basin ruled. (1) Santa Catalina Mountains, (2) Gila River, (3) San Pedro River, (4) Little Dragoon Mountains. *Olenoides pugio* time.

d.—*Earl*

Abrigo and northwestern basi
(2) Virgin Mountains.

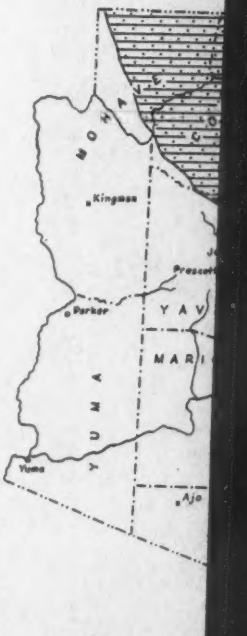
h.—*Late Mississippian*

Paradise basin ruled.



i.—*Pennsylvanian*

Pennsylvanian basin ruled. (1) Verde River, (2) Tonto Creek, (3) Lake Roosevelt, (4) Empire Mountains.



Manzano basin ruled, S
Santa Rita Mountains, (3)
tains, (5) Naco Hills.

PALEOZOIC PALEOGEOGRAPHY OF ARIZONA

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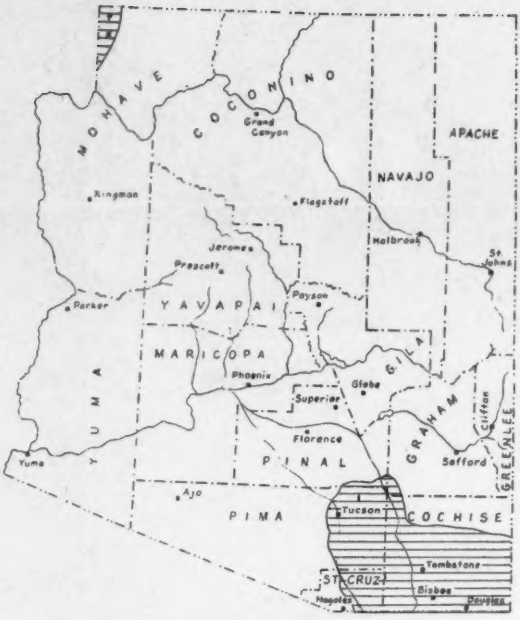
ate Middle Cambrian

led. (1) Santa Catalina Mountains, (2) Gila
er, (4) Little Dragoon Mountains. *Olenoides*



d.—*Early Late Cambrian*

Abrigo and northwestern basins ruled. (1) Santa Catalina Mountains, (2) Virgin Mountains. *Tricrepicephalus texanus* time.

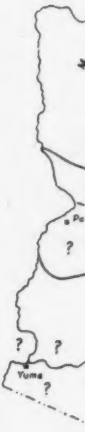
Longfellow

—Pennsylvanian

(1) Verde River, (2) Tonto Creek, (3) Lake
(4) Empire Mountains.

j.—*Permian*

Manzano basin ruled, Supai basin dotted. (1) Fort Apache, (2) Santa Rita Mountains, (3) Empire Mountains, (4) Whetstone Mountains, (5) Naco Hills.



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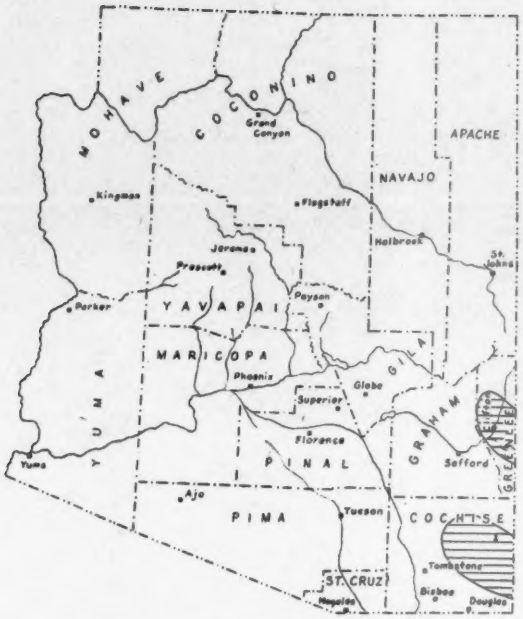
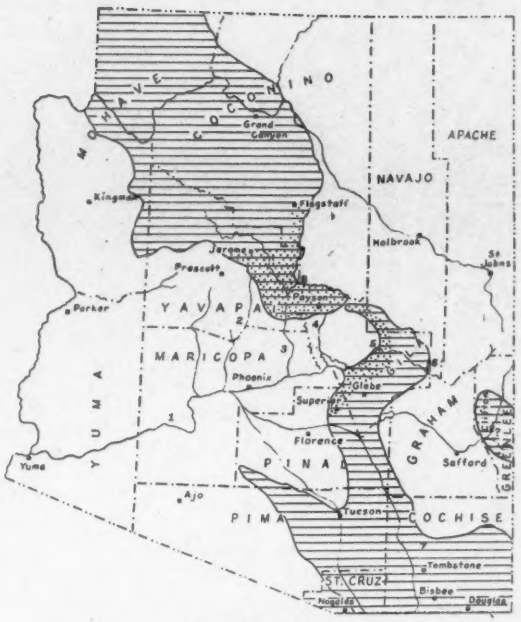


e.—*Early Ordovician*

Longfellow and Dos Cabezas basins ruled. (1) Dos Cabezas Mountains.

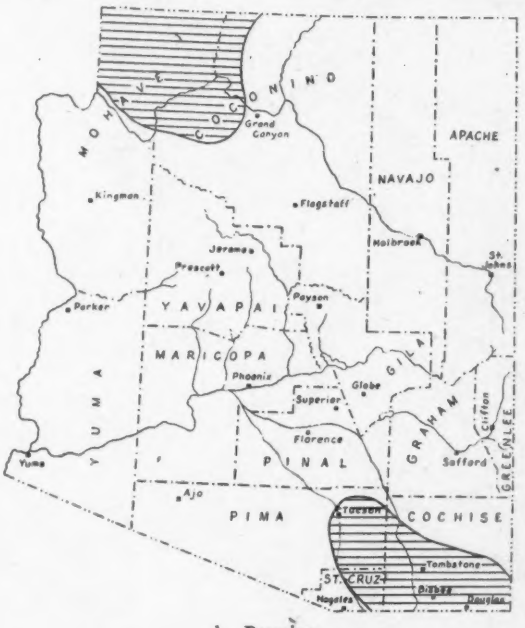


f.—*Late Devonian*

Jerome, Martin, and Morenci basins ruled, arenaceous facies with Arthrodire fishes dotted. (1) Gila River, (2) Agua Fria, (3) Verde River below its confluence with East Verde River, (4) Tonto Creek, (5) Salt River, (6) Fort Apache, (7) Little Dragoon Mountains, (8) East Verde River.

k.—*Permian*

Snyder Hill basin ruled. (1) Harquahala Mountains.



l.—*Permian*

Chiricahua and late Kaibab basins ruled. Late *Waagenoconcha montpelierensis* time.

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SHIFTINGS OF THE MISSISSIPPI RIVER IN RELATION TO GLACIATION

BY FRANK LEVERETT

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ABSTRACT

Preglacial drainage of the present Mississippi watershed was in two distinct systems. A western system headed in southern Minnesota and flowed southeastward across northeastern Iowa to Muscatine, thence southward along the present course, except for a few miles opposite the lower rapids. The eastern system from St. Paul, Minnesota, southward to the upper rapids, turned eastward at LeClaire, Iowa, and discharged through the lower part of the Illinois valley.

These drainage systems were displaced and shifted back and forth by ice sheets lying both east and west of the river. In the Nebraskan and Kansan stages disturbance by the Keewatin ice sheet initiated the cutting of a new valley from Muscatine eastward to the old eastern drainage. In the third or Illinoian stage the Labrador ice sheet covered the eastern system below Clinton, Iowa, and forced all of the drainage into a brief or temporary course through eastern Iowa outside the Illinoian ice border. When the Illinoian ice sheet melted drainage from eastern Iowa down the old valley south from Muscatine was reopened. The entire present Mississippi became established in early Wisconsin time, about 80,000 years ago. In late Wisconsin time it also carried the Glacial Lake Agassiz drainage.

INTRODUCTORY STATEMENT

In an early report, the writer (1899) discussed briefly the displacement of the Mississippi River between the upper and lower rapids by

FIGURE 1.—*Drainage map of southeastern Iowa*

Including the upper rapids at LeClaire, the lower rapids above Keokuk and features bordering the Temporary Mississippi channel in eastern Iowa. From Leverett, 1901, p. 40.

the Illinoian glaciation. The subject was little more than outlined, and did not fully interpret the length of time the river occupied the channel it opened outside the west border of the Illinoian drift; it merely stated that the river returned to the present course from Muscatine to the head of the lower rapids at Montrose when the ice sheet melted from the intervening district. That it returned rather promptly was suggested

FIGURE 2.—*The course of the Temporary Mississippi River*
From W. H. Schoewe, 1923, p. 424.

by the slight depth of the temporary channel, only 25-40 feet throughout much of its course. A long-continued stream of such volume should have done much more work, for it had a descent of more than a foot to a mile in the 75-80 miles of its course, about double the rate of the

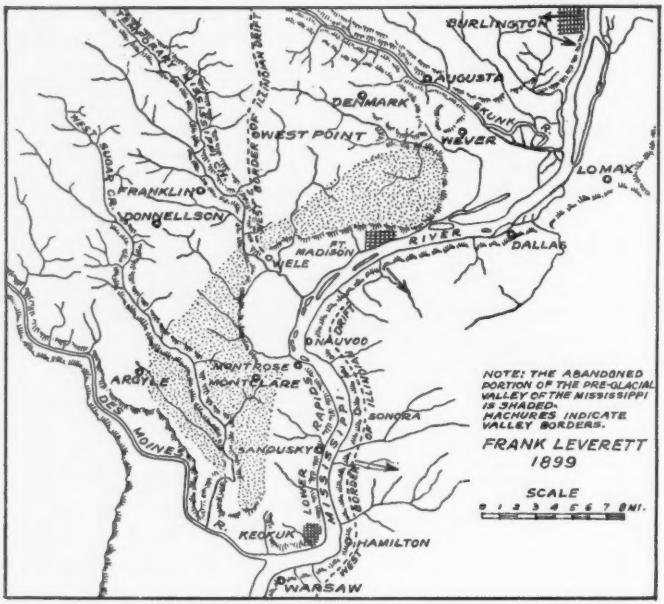

FIGURE 3.—*Sketch map of region around the lower rapids of Mississippi River*
Arrows indicate glacial striae. From Leverett, 1899, p. 468.

present river. It carried the drainage of the upper Mississippi as well as that from the Illinoian ice sheet. This temporary channel is shown in Figures 1 and 2.

Supporting evidence of an early abandonment of this channel is found in the work done by such streams as Skunk River, which turned away from this channel to open a course across the Illinoian drift and join the re-established Mississippi River. Skunk River flowed for a considerable period in a shallow valley 20-30 feet deep, but nearly 2 miles wide, which is nearly double that of the Temporary Mississippi channel, although its volume was but a small fraction of that of the Mississippi River. The shallowness was an adjustment to the level at which the re-established Mississippi started to excavate a channel in the lower

rapids, which was about 120 feet above the present level of the river at their head. Skunk River has since deepened its valley 100 feet or more in a narrow trench inside the limits of the broad, shallow valley, thus keeping pace with the lowering of the Mississippi channel across the rapids. It is of interest to note that this broad shallow valley turned south at Augusta, Iowa (Fig. 3), and headed toward the lower rapids, thus favoring the writer's interpretation of a prompt establishment of the Mississippi in the course through the rapids. Had these features been brought out more fully probably Schoewe (1924) would not have dissented from the writer's interpretation.

Figures 1 and 2, showing the course of the Mississippi in its relation to the Illinoian ice sheet, supplement each other. Figure 3 illustrates the features near the south end of the Temporary Mississippi channel, and at the lower rapids of the present river. The shallow valley of Skunk River is indicated by hachures. The filled preglacial Mississippi valley that passes around the rapids is also shown. Its size and the date of filling are discussed below.

A gumbotil of very moderate depth formed on the surface of the Illinoian drift in the early part of the somewhat lengthy period in which the river has been opening its channel through the lower rapids, a period estimated to cover about 160,000 years, and prior to the post-Sangamon loess deposition which preceded the culmination of the Labradorian-Wisconsin glaciation, estimated to have been some 80,000 years ago. The ordinary thickness of this gumbotil is 20 to 30 inches, or only one-third to one-half of the 5 feet reported by Schoewe. Very few exposures in this region have a thickness of 5 feet.

LAKE CALVIN

The name Lake Calvin was applied by Udden (1899, p. 350-356) to the lake that occupied a lowland in the lower course of Cedar River, 5 or 6 miles wide, which stood in the line of the Temporary Mississippi River. A silt deposit of this lake, exposed near Wilton, Iowa, contained the bones of a proboscidian which Calvin reported in 1874 to President Thatcher of the University of Iowa. He was unable to determine whether the bones were those of mastodon or elephant as the skull and teeth were missing. Calvin's early recognition of the silt as a lake deposit made the application of his name to the lake most fitting.

The level of Lake Calvin was determined by the altitude of the head of its spillway or outlet near Columbus Junction, which was put at 710 feet in the author's (1899) report on the Illinois Glacial Lobe.

In the 12 miles between there and Winfield it drops to 698 feet, corrected railway levels making it 5 feet less than stated in the former report. In the 25-30 miles from Winfield to Rome, on Skunk River, it drops to 670-675 feet. From there to the head of the lower rapids the

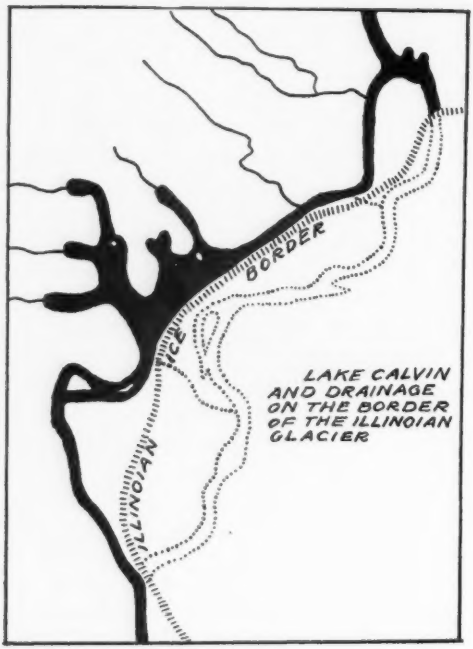

FIGURE 4.—*Lake Calvin, and drainage on the border of the Illinoian ice sheet*
From Schoewe, 1924, Plate 17.

rate is fully a foot to the mile. The channel is cut to a depth of only 25-40 feet below the bordering plain, and is less than 1½ miles wide. Its depth and breadth are so small that it has the appearance of a stream channel rather than a valley proper, the flood plain being the bordering upland.

The level of Lake Calvin was a few feet higher than the bed in the head of the outlet, for its silt deposits cover ground 720 feet above sea level on the divide between the Cedar and Wapsipinnicon rivers near Durant, Iowa. As stated by Udden, any shore lines that the lake may have formed were too weak to be traceable after so long a lapse

of time. With an altitude of 720 feet, the ponding must have extended up the Mississippi Valley far beyond the part that was covered by the Labradorian-Illinoian ice sheet, at least to the vicinity of Lake Pepin. It may there have received the drainage from the Patrician-

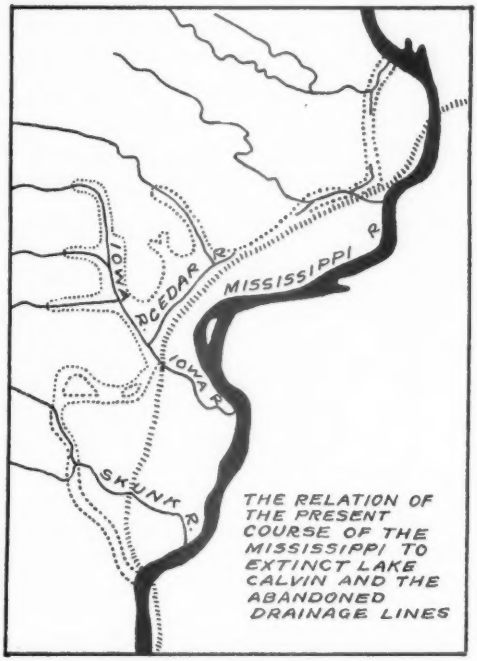
IOWA R.
CEDAR R.
MISSISSIPPI R.
IOWA R.
SKUNK R.
THE RELATION OF THE PRESENT COURSE OF THE MISSISSIPPI TO EXTINCT LAKE CALVIN AND THE ABANDONED DRAINAGE LINES

FIGURE 5.—*Relation of the Mississippi River to Lake Calvin and abandoned drainage lines*

From Schoewe, 1924, Plate 18.

Illinoian ice sheet. In this northern extension, as in Muscatine County the shore lines are probably obliterated. The depth of sediment on the bed of Lake Calvin appears also to have been very moderate, a fact that strongly favors the view that the lake was short-lived.

TIME NEEDED TO FORM THE TERMINAL MORAINE

The time required for the Illinoian ice sheet to build its terminal moraine may serve as a measure of the life of Lake Calvin, though the lake or ponded condition prevailed for a somewhat longer time—

partly before and partly after the building of the moraine, while the ice sheet advanced to and retreated from its terminal position. Where best developed the moraine is about 50 feet thick above the Kansan gumbotil and associated peat deposits remarkably preserved at its base. The building of such a moraine probably required centuries, and possibly a few thousand years. The advance to and retreat of the ice border across 20 miles intervening between the moraine and the present river seem likely to have taken a few more centuries. In the absence of more definite methods, such as it is hopd future students may work out in measuring rates of moraine building, we are only warranted in stating that at most the ice sheet occupied this district west of the Mississippi River only a very few thousand years.

WEST LIBERTY PLAIN AND OTHER FEATURES

The West Liberty Plain was described by Udden (1899). It occupies the bed of Lake Calvin in the Cedar River Lowland and at its northeast end, near Moscow, at 660 feet or about 60 feet lower than the neighboring Lake Calvin silts. He reports a downstream slope of about 3 feet per mile in Muscatine County to 615 feet at the line of Muscatine and Louisa counties.

Udden (1901) considered the terrace, 2 to 6 miles wide along the Iowa River in Louisa County, below the mouth of the Cedar River, a continuation of the West Liberty Plain. He does not clearly state the relation of this plain to Lake Calvin. However, the mention of the downstream slope of the plain and the inclusion of the Iowa River terrace in Louisa County in it, seem to indicate that he regarded it a fluvial feature, as to which the writer agrees.

Schoewe calls the West Liberty Plain of Udden the Intermediate Terrace since there is a higher terrace on Iowa River, at and below Iowa City, and also a lower one. He reports remnants of the higher terrace in the Cedar River drainage near West Liberty, Wilton, and along Cedar River above Moscow. It stands about 680 feet on Iowa River at Iowa City. It slopes about 16 inches per mile, as compared with 2 feet on the Intermediate Terrace, its altitude being 670 feet at Hills about 7 miles below Iowa City, and 660 feet 2 miles south of River Junction. It is entirely too low to find continuation over the spillway of the outlet of Lake Calvin south of Columbus Junction. The only line of continuation available appears to have been that which the Iowa River has taken. It seems to be the partially filled pre-Illinoian course of drainage. The Illinoian drift in the axis of that valley seems to have been little if any higher than 650 feet. A

detailed study along the borders of the present Iowa River valley below the mouth of the Cedar River may reveal remnants of the high terrace, the broad terrace in the valley being, as suggested by Udden, a continuation of the West Liberty Plain (Schoewe's Intermediate Terrace).

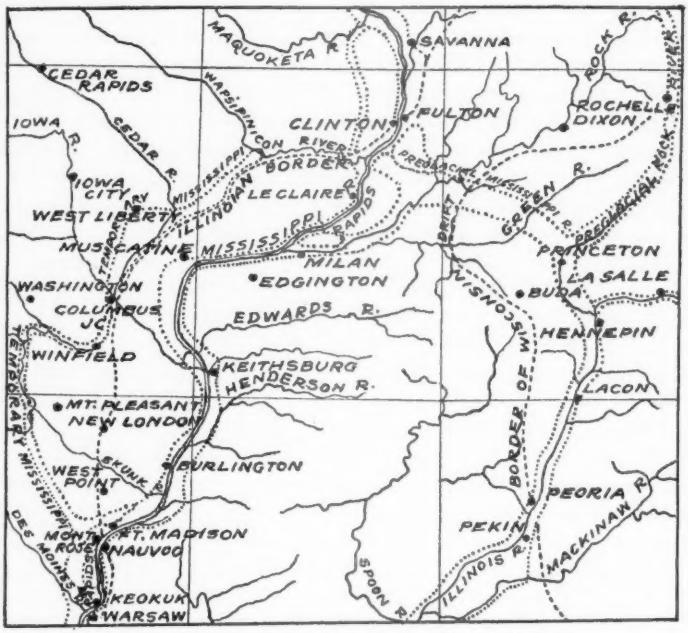

FIGURE 6.—*Map showing shiftings of Mississippi River*

Also the preglacial Rock River valley, the position of the Illinoian and Wisconsin ice borders, and the upper and lower rapids. From Leverett, 1921, p. 625.

The Intermediate Terrace declines from about 650 feet at Iowa City to 615 feet at the Louisa County line, and to 606 feet at Fredonia, opposite the head of the Lake Calvin spillway (710 feet) near Columbus Junction, thus being fully 100 feet below that lake level. The High Terrace and the Intermediate Terrace are sandy, and not silty like the Lake Calvin deposits, except in a side valley near Wilton (Schoewe (1924), Figs. 25, 26, 27). In their make-up as well as in their downstream slope, they are fluvial rather than lacustrine. The Low Terrace, on the Iowa River below Iowa City, is of coarser material and is regarded by Schoewe as fluvial. The writer sees no reason for classing

the others as lacustrine, or for continuing the life of Lake Calvin beyond the time its outlet through the Temporary Mississippi channel was in operation.

PRE-ILLINOIAN SHIFTINGS WITHIN THE MISSISSIPPI DRAINAGE BASIN

At the opening of the Glacial Epoch the upper Mississippi Valley above LeClaire, Iowa, and the valley of the present Mississippi below Muscatine, belonged to independent drainage systems. The upper Mississippi, and the Rock River drainage, flowed southeastward through the Green River Lowland to the Illinois at Hennepin, and thence down the Illinois Valley to its mouth (Fig. 6). Another drainage system which appears to have headed in southern Minnesota flowed southeastward across northeastern Iowa to Muscatine. Its course and position have been only rudely determined from the well data of the region. The wells in its valley penetrate 200-300 feet of drift, while those outside the valley and main tributaries generally enter rock at less than 100 feet. More detailed collection and sifting of well records is needed to determine the course of the river and the limits of its drainage basin. A suitable name for it might then be given.

Below Muscatine, with the exception of a few miles near the lower rapids, this old valley is followed by the present Mississippi to the mouth of the Illinois River, where it connects with the eastern drainage from the upper Mississippi Valley. This combined stream, further augmented a few miles below by the Missouri River, utilized the large preglacial valley as far as Thebes, Illinois. It there partially abandoned the old valley and cut through a narrow peninsula to join the Ohio River. The valley is still utilized in flood stages below Thebes.

Warren (1878) reported the old or preglacial course around the lower rapids as indicated in Figure 3. Its valley is about 6 miles wide. Gordon (1890) reported a well record at Mont Clare, Iowa, which is within this abandoned part of the old valley. This shows its bed to be 126 feet lower than that of the present stream at the head of the rapids, or 374 feet above sea level. The filling of 305 feet is mainly Nebraskan drift. The overlying Kansan drift is only about 50-60 feet thick in that vicinity. The post-Kansan loess is but 6-12 feet thick. The Nebraskan drift is thus about 240 feet thick. The Paper Company well at Fort Madison, 11 miles up the valley, on a low terrace penetrated 145 feet of filling, mainly Nebraskan till, and entered rock at 379 feet above sea level. This is 123 feet below the low water level of the present stream at that point. The filling was sufficient at least by Kansan time, and very likely by Nebraskan time to permanently dislodge the stream from its preglacial course.

This seems a fitting time to give attention to the effect of the Kansan glaciation on drainage in part of the Mississippi Valley. It presents an interpretation held for some time but not hitherto published, and which is only vaguely expressed elsewhere in the literature.

The Kansan glaciation by the Keewatin ice sheet extended eastward several miles beyond the Mississippi River between Hannibal, Missouri, and Fort Madison, Iowa, and at least to the west bluff of the valley for 30 miles farther north, to Kingston, Iowa, where glacial striae trend southeastward. Striated ledges with this bearing are also found 10 miles east of the east bluff of the valley in the latitude of Quincy, Illinois. The position and bearing of the striated ledges near Hamilton and East Fort Madison, Illinois, and on the bluff 2 miles south of Burlington, Iowa, are indicated on Figure 3, as well as a later set of westward bearing striae at Burlington, referable to the Illinoian glaciation.

Evidently the displacement of drainage in this part of the Mississippi Valley was complete in Kansan time. Probably at this time some of the work of opening the valley eastward from Muscatine was enacted. Work probably started there in the preceding glacial stage, for the Nebraskan glaciation was similarly related to that part of the Mississippi Valley. However, that is an open question depending on whether the Nebraskan drift filling was high enough to prevent the Mississippi from returning to its pre-glacial course west of the lower rapids. In any case, the water from the melting Nebraskan ice sheet probably found eastward discharge there. How long the post Nebraskan drainage may have utilized this line is uncertain. As to the post-Kansan drainage through this line, very probably it persisted in the eastward course until the Illinoian ice invasion from the Labradorian Center; also the drainage of Iowa River may have utilized it. As to Skunk River, its present lower course is post-Illinoian. The pre-Illinoian course seems likely to have continued eastward from Brighton in southern Washington County, and it may have joined Iowa River. It can be affirmed with certainty that the Des Moines River basin drained southward in its present course after the Kansan glaciation.

The new section of the Mississippi Valley under consideration may have been much smaller than now when its eastward-flowing drainage gave place to the great stream now flowing westward. The size of the valley is well shown in the Edgington and Milan topographic maps of the U. S. Geological Survey. It averages about 1½ miles in width and has steep bluffs 150-200 feet high. Standing as it does within the district invaded by the Labradorian ice sheet in the Illinoian stage the valley was probably partially filled with Illinoian drift. On the border-

ing uplands this drift is only 20-30 feet deep. Unless the valley received a greater amount, the Mississippi River would have required only a relatively short time to remove it. The preglacial col or drainage divide in this valley was thought by Udden to have been near Fairport, Iowa; but the pointing of the present drainage on the borders of the channel seems to place it near Montpelier, 5 miles farther east. The upland there is fully as prominent as near Fairport.

CHANNELING AT THE UPPER RAPIDS

The author (Leverett, 1899) left open the question of the date of establishment of the Mississippi River in its course through the upper or LeClaire rapids. The work accomplished in eroding its trench is very slight compared with that involved in the trench at the lower rapids. The rock excavation is restricted to a short section, only 2 to 3 miles long, and nowhere more than 75 feet deep, the remainder being in loess and drift. However, the energies of the Mississippi River have been divided among several lines of discharge in that vicinity making it difficult to evaluate this part of its work. The broad Meredosia Slough, as may be seen by reference to the Cordova topographic map (Leverett, 1899, Pl. 18) is below 600 feet and is within the reach of present high stages of the Mississippi. At flood the stream joins Rock River and passes down its valley to the Mississippi, detouring the east side of the upland east of the upper rapids trench. The Cattail Slough makes a similar detour of the Mississippi flood waters a few miles farther north. It now stands above their reach because of accumulated peat and muck beds.

The Mississippi River seems to have divided its waters here for a considerable period in post-Illinoian time. Drainage may have been southeastward to the Illinois Valley through the Green River Basin (Fig. 6), about to the time of the culmination of the Wisconsin glaciation, when the strong moraines of early Wisconsin drift were formed. These have completely blocked the drainage from the upper Mississippi River to the Illinois Valley. It is estimated that this early Wisconsin culmination dates back about 80,000 years, or about half post-Illinoian time. The work at the upper rapids may thus be mainly within Wisconsin and post-Wisconsin time. Stream piracy by small tributaries of the Mississippi may have figured in opening this line of drainage, which has the advantage of being much shorter than the broad valleys that detour around the rapids.

This pre-Wisconsin drainage of the upper Mississippi to the Illinois Valley seems to require that the waters between the Illinoian and

Wisconsin glaciations that cut the lower rapids trench should involve mainly the district below Muscatine. The Cedar-Iowa drainage would then become the headwater line, with Skunk River the chief tributary, there being only small streams on the eastern or Illinois side. This interpretation may help to account for the slowness of the opening of the lower rapids trench.

VALLEY FILLING BELOW THE LOWER RAPIDS

The Mississippi Valley was filled at the lower end of the rapids during and after the Illinoian glaciation to about 90 feet above the present river, forming a terrace which at Warsaw, Illinois, stands a little above 560 feet. The Des Moines River built up its valley to a similar height at the "Yellow Banks" near its mouth. A section there, published by Gordon (1895, p. 243) is of especial interest in showing a succession of deposits, separated by two beds of black earth.

"YELLOW BANKS" SECTION

	Feet	Inches
(1) Clay, yellow, pebbleless (loess)	5	
(2) Silt, drab, pebbleless (loess)		15
(3) Earth, black, with a few small pebbles, apparently a flood-plain deposit	12	
(4) Clay, yellowish, local		6
(5) Sand, with a few small pebbles, with a layer of boulders at base	20–25	
(6) Earth, black, with yellow streaks, apparently an old flood-plain deposit	5–6	
(7) Gravel, with some sand, pebbles 2 inches or less in diameter	20	
(8) Clay, blue till, exposed to level of railway	15	
Total, about	80	

The upper bed of black earth, being at the base of the loess, seems to fall in Sangamon time and correlate with the Illinoian gumbotil. The thickness of 12 feet indicates a considerable time interval in harmony with the time involved in the weathering exhibited by the Illinoian gumbotil. The underlying pebbly sand, adjacent to that in the Warsaw terrace, connects with the filling from the excavation of the lower rapids, made at the mouth of the Des Moines. The lower black earth and the gravel below it seem to be pre-Illinoian, and probably post-Kansan, while the blue till at the base is either Kansan or Nebraskan. The low altitude rather favors its reference to the Nebraskan glaciation, the top of the deposit being only about 500 feet above sea level. The surface of the Kansan drift in the upland back of the Yellow Banks, as shown by the Kahoka, Missouri-Iowa topographic map, is about 660 feet. Allowing 60 feet for the Kansan drift,

the Nebraskan drift should reach about 600 feet, or 100 feet above its exposure at Yellow Banks.

At Keokuk, and at points above, boulders are conspicuous 50 feet or more above the river. (*See* Leverett 1899, Pl. 7.) These probably are a concentrate from erosion of the Illinoian till where its edge was invaded in the development of the gorge. At Keokuk the boulders are under a sandy gravel the level of whose surface seems to correlate with the filling at Warsaw. Below Warsaw the altitude of the valley filling decreases much more rapidly than the fall of the present stream. Opposite Hannibal the filling at the mouth of Hadley Creek is only 35 feet above the river. The river falls only 25 feet in this interval while the terrace falls 90 feet.

PERMANENT ESTABLISHMENT OF THE MISSISSIPPI RIVER

The Mississippi River is in a new course above St. Paul. Winchell (1888), Grant (1890), and Sardeson (1908, 1916), measured post-Wisconsin time by recession of the waterfalls. The deeply-trenched valley at St. Paul is only a minor tributary of the preglacial upper Mississippi, the main line being farther west, beneath Lake Minnetonka. Its rock bed there is about 500 feet above sea level, or 180 feet below the level of the present river at St. Paul. Its course was southeastward through Dakota County and the St. Paul tributary entered it near Hastings. From Hastings, as far south as Clinton, Iowa, the river has a bedrock floor averaging fully 100 feet below the present stream. The width of the valley is nowhere less than 2 miles, and it averages about twice that width. As it is below the average width in the vicinity of Dubuque, Hershey (1897) suggested that the preglacial river headed only a short distance above that city. This view does not seem to find support in the slope of the rock bed, since no part of the floor slopes distinctly in the reverse direction as one approaches the head of the present stream. This continuous deep channel also seems to oppose a view held with some persistence by Trowbridge (1921) that this part of the Mississippi Valley is post-Nebraskan.

Just below St. Paul, at the Wisconsin drift border, the valley was remarkably aggraded with outwash gravel up to 940-960 feet—260-280 feet above the river. From there the surface of the outwash terrace descends fully 100 feet in the first 20 miles. It is only 790 feet above sea level at Red Wing and 125 feet above the river. At the mouth of Wisconsin River it is 75 feet above the Mississippi River, and 60 feet at Bagley, Wisconsin, 12 miles below. It then maintains a height of 50-60 feet above the river clear down to the head of the lower rapids.

The excavation of the lower rapids gorge may have been only down to this level when the Wisconsin filling reached there, thus helping to determine the height of the filling.

In the waning part of the Wisconsin stage the outlet of the great Glacial Lake Agassiz the "River Warren", carried a flood that swept away much of the Wisconsin filling of sandy gravel, and brought the valley about to its present condition. The flood widened the valley and cut back and greatly steepened the projecting headlands. The volume must have been far greater than in any of the modern high water stages. The lake lasted but a few thousand years and ended soon after its outlet was diverted to Hudson Bay. The entire life of the lake probably was not more than 10,000 years, its final 1000 years being tributary to Hudson Bay.

SUMMARY

Summarizing the succession of drainage shiftings, it appears that the two preglacial drainage systems became united by an eastward diversion of the western system by the Keewatin ice sheet in the Nebraskan and Kansan stages into the valley of the eastern system. At the culmination of the glacial stages it was largely ice sheet drainage, but later it may have become the course of interglacial drainage from the western area as far down as the lower rapids until the Illinoian glacial stage, in which case its size was similar to that of the present Mississippi.

In the Illinoian stage the valley of the eastern system was covered by the ice sheet below Clinton, Iowa, and also the valley of the western system from Muscatine south to the lower rapids below Fort Madison. So the river was forced to open a new valley outside the border of that ice sheet. This carried the upper Mississippi from above the Illinoian border, as well as the discharge from the melting ice sheet, so its volume somewhat exceeded that of the present river. This was only a temporary course, restricted to the time the valley from Muscatine to Fort Madison was ice-covered. Then, when the Illinoian ice border had shrunk eastward past the Illinois River below Hennepin, the eastern system may have returned to its old course. In that case, the western system would have embraced only the drainage from eastern Iowa and the western edge of Illinois, and would have had a watershed only a small fraction of that of the present Mississippi. This may have been operating throughout the interglacial (Sangamon) stage between the Illinoian and Wisconsin glacial stages, a period of perhaps 100,000 years.

It was in the Wisconsin glacial stage that the Mississippi became permanently established in its present course and drainage area. However, for a few thousand years in the waning part of the Wisconsin stage it received the discharge from the Glacial Lake Agassiz, which apparently more than doubled the flow through the valley.

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"PSEUDOMIGMATITE" IN THE PIEDMONT OF MARYLAND

BY RANDOLPH W. CHAPMAN

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ABSTRACT

The Laurel gneiss lies in the vicinity of Laurel, Maryland, about 17 miles southwest of Baltimore. A cover of Cretaceous gravels obscures all but about 10 square miles. On the north the Laurel gneiss is bounded by Guilford granite and the Wissahickon formation, on the northeast by gabbro, on the east by Cretaceous gravels, on the south by gneiss, and on the west by the Wissahickon formation.

The Laurel gneiss is gray, medium-grained to fine-grained, and faintly foliated. Chief minerals are quartz, oligoclase, biotite, and muscovite. Remnants of schist and quartzite from the Wissahickon formation are common.

Although the Laurel gneiss looks like a true migmatite, it is believed to have originated by the granulation, flowage, and recrystallization of the Wissahickon formation under conditions of stress, high temperature, and abundant water. This is suggested by: (1) the mineralogical and chemical similarity of the two rocks, (2) the granulated and recrystallized texture of the gneiss, (3) the abundance of schist and quartzite remnants which do not show granulation or recrystallization, and (4) the complete gradation from Laurel gneiss into the Wissahickon formation across contacts. The writer thus suggests the name "pseudomigmatite" for this body.

The Laurel gneiss has most of the features of a plutonic igneous rock but it does not show intrusive contacts. It is thought, however, that if conditions of metamorphism had been more intense it might have developed even these. The origin of the Laurel gneiss appears to be different from that of other igneous-looking rocks.

PROBLEM

In the Spring of 1940 the new geological map of Howard County, Maryland, prepared by Ernst Cloos and Carl Broedel, was published by the Maryland Geological Survey. In the southeast corner of this area is a body of rock which the authors named the "Laurel migmatite".

Cloos believed that it originated by granitization. He pictured the Wissahickon formation as having been thoroughly penetrated and assimilated by intrusions of granitic magma. His interpretation was based mainly upon field studies, however, and, since the "migmatite" has many of the megascopic aspects of an igneous rock with inclusions of schist and quartzite, it is not strange that he should have adopted this view.

Shortly afterward, the present writer undertook a detailed study of the "Laurel migmatite". The areal distribution of the rock was mapped, and the structure and petrography were investigated. As a result, a new theory of origin is advanced. Briefly, the writer believes that the "Laurel migmatite" is not a migmatite but a metamorphic rock which has been produced by the granulation, flowage, and recrystallization of the Wissahickon formation, under conditions of stress, high temperature, and abundant water. The term "migmatite" is, therefore, no longer applicable. However, since the rock has the field appearance of a true migmatite, the writer believes that the term "pseudomigmatite" is more appropriate. In this study the name "Laurel gneiss" is used for this body in preference to a name with more genetic significance.

ACKNOWLEDGMENTS

The writer is especially indebted to Ernst Cloos for his generous assistance in the preparation of this report. He has prepared the section dealing with fabric analyses. Thanks are also due Marland P. Billings and Carleton A. Chapman for reading and criticising the manuscript. Mark H. Secrist, Judson L. Anderson, and James Bell assisted the writer in the preparation of photographs, and this assistance is gratefully acknowledged. George Tunell was kind enough to determine the index of refraction of a specimen of garnet. Through the kindness of Frank L. Hess, spectrographic analyses of two specimens of biotite were made by Morris Slavin at the United States Bureau of Mines. Thin sections for petrographic study were obtained through the generosity of the Maryland Geological Survey.

GEOLOGICAL RELATIONS

The Laurel gneiss lies in the vicinity of Laurel, Maryland, about 17 miles southwest of Baltimore. The body has a length of about 16 miles and extends almost into the city of Washington. The width is uncertain, but it is at least 3 miles. Due to the Cretaceous cover, only about 10 square miles of the gneiss are actually exposed (Pl. 1).

The Laurel gneiss is bounded on the north by the Wissahickon formation and the Guilford "granite" (granodiorite), dikes of which cut the

gneiss in many places. To the northeast lies a body of coarse, dark, foliated gabbro whose contact with the Laurel gneiss has never been seen. However, near locality 1 (Pl. 1) the gneiss is cut by a dikelike intrusion which resembles the gabbro petrographically. This suggests that the gabbro is younger. On the east, the Cretaceous sands and gravels of the Coastal Plain lap upon the gneiss and obscure much of its areal extent. The Wissahickon formation bounds the Laurel gneiss on the west and grades into it across a transitional boundary. The structure of these two rocks is closely conformable as may be seen from the geological map (Pl. 1).

A medium-grained, gneissic rock (probably the Sykesville granite of Cloos and Broedel) terminates the Laurel gneiss on the southwest. This gneiss has, locally, some of the features of the Laurel gneiss such as quartzite fragments (inclusions) and quartz lumps, but it differs by being strongly foliated. It is, therefore, probably older than the Laurel gneiss. The boundary between the two gneisses is difficult to locate owing to a lack of good exposures in critical areas and to the similarity of the two rocks. A dashed boundary, located with the aid of outcrops farther to the north, is, therefore, shown on the geological map (Pl. 1).

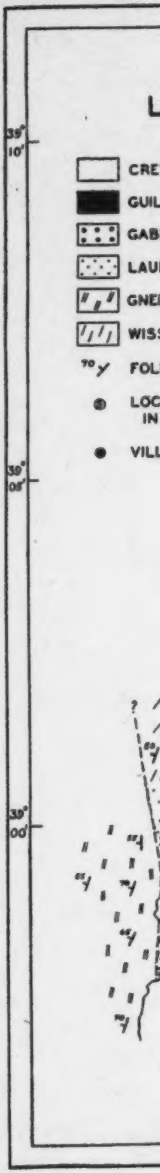
Field localities where specimens were collected, photographs taken, or special features observed are numbered on the geological map (Pl. 1). Since the Laurel gneiss is best exposed in the vicinity of Laurel, most references are made to this locality. However, excellent exposures may also be seen along the two streams flowing southeast from localities 39 and 56.

DESCRIPTION OF THE LAUREL GNEISS

GENERAL FEATURES

The Laurel gneiss is a fine-grained, porphyroblastic rock consisting principally of quartz, plagioclase, muscovite, and biotite. It is light-gray except for those portions which surround mica-rich segregations. (*See* section on segregations.) Where weathered it is yellowish-brown. Megascopically the rock appears fairly uniform in grain size; individual grains range between ¼ millimeter and 1 millimeter in diameter. Some porphyroblasts, however, may reach 2 or 3 millimeters. The texture of the gneiss is granular or sugary.

Structure in the gneiss is not readily apparent everywhere, but many of the fine mica flakes have a rough planar orientation. This produces a foliation which is uniform in strike and dip and conformable with the regional structure. It trends somewhat east of north and dips steeply west. Locally, as for instance at a point 1000 feet northwest of locality 22 (Pl. 1), the foliation is thrown into a series of small but very complex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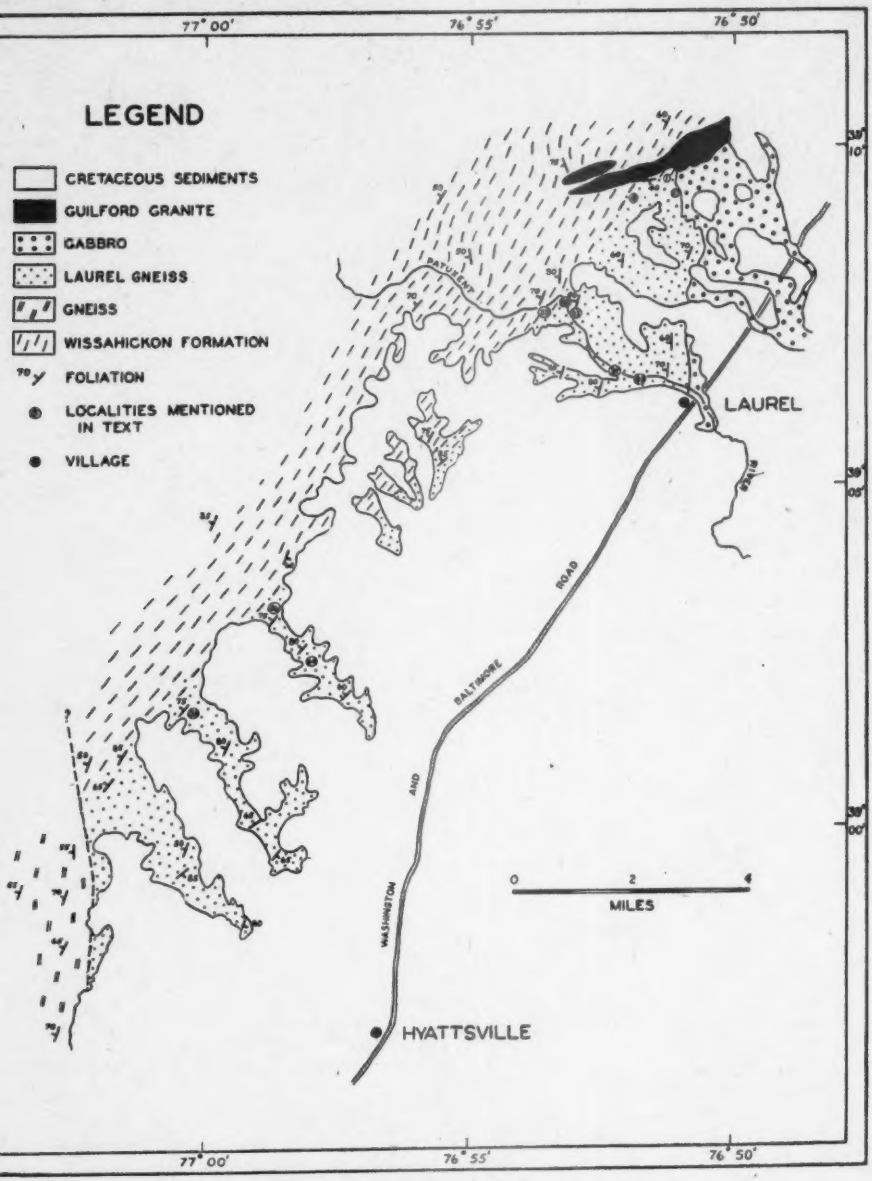

GEOLOGICAL MAP OF THE LAUREL GNEISS, MARYLAND

folds. In other places it is irregular or wavy. Elsewhere the foliation wraps around remnants of older quartzite or is distorted by them. All these facts suggest strongly that the Laurel gneiss has actually flowed as a plastic mass.

In many places the rock is streaked or banded by long, irregular or lenslike segregations composed principally of coarse biotite and muscovite crystals with diverse orientations. These masses greatly accentuate the foliation of the gneiss since they are streaked out parallel to the latter.

The general appearance of the normal Laurel gneiss in the field is that of an intrusive rock rather than of a metamorphic type. Where remnants of older schist and quartzite are abundant, however, it has more the appearance of a hybrid rock.

MICROSCOPIC FEATURES

Texture.—Under the microscope the Laurel gneiss is holocrystalline, inequigranular, and allotriomorphic (Pl. 6, fig. 2). Quartz and plagioclase compose much of the rock. The plagioclase crystals are generally equidimensional, but the quartz grains show an elongation parallel to the foliation. Adjacent grains of these minerals are generally interlocked by "sutured" boundaries which are characteristic of recrystallized rocks. Small biotite and muscovite crystals, scattered throughout the groundmass, are roughly oriented, giving the rock its poor foliation. Coarser crystals of biotite and muscovite occur as porphyroblasts cutting the foliation at all angles.

The minerals are distributed unevenly throughout the rock; some areas consist mainly of quartz and plagioclase, and others are very rich in mica. Accordingly, it is impossible to give one mode which will be exactly representative of all specimens. Table 1 shows the modes of several specimens. It will be noted that the same minerals are present in each specimen although they vary in amounts.

Quartz.—Clear, glassy quartz is the most abundant mineral. It occurs as anhedral grains which are elongated parallel to the foliation. The length of the grains (¼ to 1 millimeter) is generally from two to three times the width. Individual crystals are interlocked with surrounding grains of quartz and plagioclase by sutured boundaries. Undulatory extinction is common.

In the transition zone between the Laurel gneiss and the Wissahickon formation large crystals of quartz show special evidence of straining and deformation. Many of these have been broken into fragments. In some instances the fragments have been displaced only slightly. In other cases, however, they are so greatly displaced that evidence of their original relationships is almost lost (Pl. 6, fig. 4).

Plagioclase.—Most of the plagioclase crystals are anhedral and equidimensional. Some appear to be interstitial between quartz grains, suggesting that they crystallized later than the quartz. Most of the plagioclase is calcic oligoclase ($Ab_{72}An_{28}$). One or two sections show a slightly more calcic oligoclase, whereas one contains andesine ($Ab_{63}An_{37}$). Many of the grains are twinned polysynthetically. Straining is much less common than in the quartz. On the whole the oligoclase is clear, but a few grains show alteration to sericite.

Micas.—Muscovite and olive-brown biotite are both present in the gneiss although muscovite is somewhat more abundant. There are two generations of each mica.

The small crystals of the younger generation, scattered more or less evenly throughout the rock groundmass, are clear and free from inclusions. They have a rough planar orientation. Large porphyroblasts of the later generation show diverse orientations. The biotite porphyroblasts have frayed edges and contain abundant poikiloblastic inclusions of quartz, plagioclase, magnetite, and zircon. In some, inclusions are so abundant that the porphyroblasts are mere skeletons. The muscovite porphyroblasts have more regular boundaries than the biotite porphyroblasts. They contain inclusions of quartz, plagioclase, garnet, magnetite, apatite, and zircon. Many muscovite and biotite porphyroblasts are intergrown with one another parallel to cleavage faces suggesting simultaneous crystallization. Both the muscovite and biotite porphyroblasts show wavy extinction.

The optical properties of the muscovite are as follows: $\beta = 1.605$, $\gamma = 1.610$; optically negative; $2V = 40°$; dispersion $r<v$, weak. These properties are similar to those given by Barth (1936, p. 780-781) for a muscovite from Dutchess County, New York, and suggest that the chemical composition is approximately the same. Barth's analysis of the Dutchess County muscovite is as follows:

SiO_2	46.31
TiO_2	0.79
Al_2O_3	38.52
Fe_2O_3	tr
FeO	....
MgO	0.18
CaO	0.46
Na_2O	1.56
K_2O	8.03
Li_2O	tr
H_2O	4.46
Total	100.31

Since the optical properties of the muscovite porphyroblasts are the same as those of the oriented muscovite crystals in the groundmass, their chemical composition is doubtless the same.

Optical study of the biotite shows the following properties: $\alpha = 1.590$, $\beta = 1.641$, $\gamma = 1.641$; optically negative; $2V =$ ca. 3°; dispersion $r<v$, medium to strong; pleochroism, X = pale yellow-brown, Y = olive-brown, Z = olive-brown, $X<Y = Z$. However, the large porphyroblastic biotites which have been strained have an optic angle ranging from 10° to 15°. According to Winchell (1933, p. 274, Fig. 200) the above optical properties indicate common biotite with FeO:MgO = 55:45. Both the biotite porphyroblasts and the fine, oriented biotites have the same optical properties so their chemical compositions are probably the same. A few slides show biotite altering to chlorite.

Garnet.—Pink garnet occurs chiefly as large, rounded crystals. In many slides the fine, oriented micas of the groundmass were seen to bend around the garnet crystals, proving that in general the garnets are older. In one or two instances, however, garnets appeared to cut across quartz, plagioclase, and mica. Poikiloblastic inclusions of quartz, magnetite, biotite, and muscovite have been found in the garnet. The index of refraction is 1.815± .010. According to Eskola's diagram (1921, p. 9) this indicates almandine with the following composition:

Mol. per cent	
87	$Fe_3Al_2Si_3O_{12}$
5	$Mg_3Al_2Si_3O_{12}$
8	$Ca_3Al_2Si_3O_{12}$

Tourmaline.—Tourmaline is an accessory in the Laurel gneiss, but optical data were obtained in order that this mineral could be compared with that in the Wissahickon schist. These optical properties are as follows: $\varepsilon = 1.631$ (pale yellow-pink), $\omega = 1.661$ (deep blue-green). Larsen and Berman (1934, p. 247) describe a tourmaline with similar optical properties which has the following chemical composition:

$$(Na, Ca)\ R_3\ Al_6\ B_3\ Si_6\ O_{29}\ (OH)_2$$

in which $Na:Ca = 12:1$, and $R = Mg:Fe'':Fe''' = 2:3:5$.

Accessory minerals.—In addition to the tourmaline, accessory minerals consist of magnetite, zircon, apatite, chlorite, and sericite.

REMNANTS

Introduction.—The Laurel gneiss contains many remnants of the Wissahickon formation. The reason for calling these remnants rather than inclusions will be realized later. They are of three different types: (1) coarse biotite-muscovite schist, (2) fine biotite-muscovite quartzite, and (3) quartz lumps.

Schist.—The remnants of coarse biotite-muscovite schist may be studied best at locality 22 and along the Patuxent River to the northwest (Pl. 1). They occur as bands or irregular masses with ragged boundaries and range from a few inches up to 10 feet or more in width. Most of the remnants are oriented with their long axes parallel to the foliation. The number of remnants differs greatly. In some of the outcrops northwest of locality 22 (Pl. 1) remnants comprise about 50 per cent of the rock. In other outcrops in this vicinity the remnants are few and scattered. Some of the exposures around locality 22 (Pl. 1) are so rich in remnants that the rock appears more like the Wissahickon schist than it does like the Laurel gneiss.

The schist remnants are coarse, recrystallized, porphyroblastic, and distinctly foliated. The minerals of the groundmass have an average diameter of about 1 millimeter, but because of abundant porphyroblasts the rock appears much coarser. The porphyroblasts consist chiefly of large muscovite crystals with diameters of from 5 to 15 millimeters which give the remnants a brilliant spangled appearance. Individual crystals of tourmaline are conspicuous, some attaining a length of nearly 2 centimeters. Other porphyroblasts are red garnets (diameter 1 to 2 millimeters) and biotite (diameter 2 to 3 millimeters).

Under the microscope the schist remnants are inequigranular and decidedly schistose. They consist of muscovite-rich layers which alternate with layers of quartz and calcic oligoclase. The muscovite crystals in the groundmass show an almost perfect planar orientation. The coarse porphyroblasts of muscovite, garnet, olive-brown biotite, and

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blue-green tourmaline cut directly across the structure of the groundmass indicating clearly that they are younger.

There can be little doubt that these remnants represent fragments of Wissahickon schist. In the first place, the textures, modes, and sequences of crystallization of the two rock types are the same. Furthermore, optical data suggest that the minerals of the remnants have the same chemical composition as those of the Wissahickon schist.

Quartzite.—Light gray, fine-grained quartzite remnants also are abundant in the Laurel gneiss. These may be observed best at localities 20, 22, and 35 (Pl. 1). Many are elongate with a length of from 1 inch to 30 inches and a width of approximately one-third their length. Others are more equidimensional or irregular with diameters of from 1 to 12 inches. Most of the quartzite remnants show fine banding. In those which are elongate the banding is generally parallel to their longest dimension, which in turn is almost always parallel to the foliation of the gneiss. Many of the equidimensional or irregular ones, however, lie with the fine banding of the quartzite at an angle to the foliation. Several of these may be seen at locality 35 (Pl. 1). It is important to note that near the remnants the foliation in the Laurel gneiss commonly bends around parallel to the boundaries. This indicates clearly that the foliation is younger than the remnants.

The contact relations between a fine-grained quartzite remnant and the surrounding Laurel gneiss are also shown at locality 11 (Fig. 1). Here an irregular quartzite block, 6 inches across, and with no visible foliation, is inclosed in the gneiss. For the most part, the foliation of the gneiss is parallel to the boundaries of the block. Near the lower end of the quartzite block, the foliation of the gneiss has been thrown into a series of small folds. It is clear, therefore, that the foliation of the Laurel gneiss is younger than the quartzite remnant since it has been deformed by the latter.

One quartzite remnant at locality 35 (Pl. 2, fig. 2) has been faulted and displaced about 4 inches. The foliation of the Laurel gneiss, which wraps around the remnant, has also been displaced, but the break has been sealed by recrystallization.

The quartzite composing the remnants is a light-gray, sugary rock with a grain size of from $\frac{1}{8}$ to $\frac{1}{2}$ millimeter. Microscopically it is equigranular consisting mainly of interlocking grains of strained quartz with some plagioclase. Minute crystals of olive-brown biotite and somewhat larger grains of muscovite show a subparallel orientation. A few large muscovite, biotite, and garnet crystals form porphyroblasts.

The remnants described above are similar both texturally and min-

eralogically to the fine-grained quartzite layers interbedded with the Wissahickon schist. Modes of two of these quartzite layers (20B and 25) are shown in Table 1. This similarity leaves little doubt the fine-grained masses in the Laurel gneiss are remnants of Wissahickon quartzite.

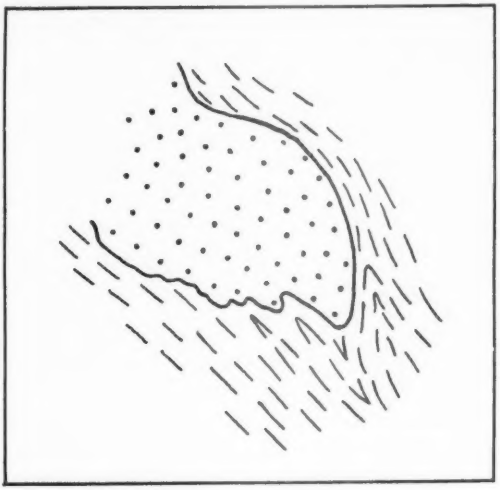


FIGURE 1.—*Quartzite remnant in Laurel gneiss*

Sketch, drawn at locality 11 (Pl. 1), showing the effect of a Wissahickon quartzite remnant (dotted) on the foliation of the gneiss (dashed). Width of quartzite remnant about 6 inches.

Quartz lumps.—Small irregular lumps of glassy or milky quartz may be studied best at locality 20 and along the Patuxent River to the west (Pl. 1). Their diameter ranges from a quarter of an inch to 4 inches; the majority have a diameter of about 1 inch. Some of the quartz lumps, are elongate or lenslike with the longer axis parallel to the foliation. Others, however, are spheroidal or irregular and show no special orientation. The lumps stand out on weathered surfaces because they are more resistant than the surrounding gneiss.

Under the microscope each lump is seen to consist of a group of large, highly strained quartz grains (Pl. 6, fig. 3) completely interlocked with one another by sutured boundaries, thus indicating thorough recrystallization. The boundary of the lump is likewise sutured as a result of intercrystallization of the lump with the crystals of the surrounding gneiss.

The quartz lumps are most likely recrystallized remnants of thin quartz layers and veins which occur in the Wissahickon formation.

SEGREGATIONS

Dark segregations are common in the Laurel gneiss. These consist of accumulations of large porphyroblasts of biotite and muscovite with occasional garnets. Individual crystals vary from 1 to 5 millimeters in diameter. In general, they show random orientations.

The segregations are of two different types. Most common are mica masses pulled out into dark, irregular streaks or lenses more or less parallel to the foliation of the gneiss. These streaks have an average width of from ⅛ to 1 inch, and a length of from 1 inch to more than 1 foot. Where the streaks are abundant the gneiss is strongly banded.

The second type of segregation is best demonstrated at locality 20 and at a point on the Patuxent River 500 feet west (Pl. 1). At these localities the masses of unoriented mica are rounded or irregular, and their diameter varies from ¼ inch to several inches. The dark segregations are surrounded by a zone of cream-colored material, ½ inch to 4 inches wide (Pl. 4, figs. 1, 2; Pl. 5, figs. 1, 2). Thin sections show that this light-colored zone differs from the normal gray gneiss only because it is relatively deficient in both biotite and muscovite. It is quite evident, therefore, that this type of segregation has formed at the expense of the micas in the immediately surrounding gneiss. Its origin will be considered later.

WISSAHICKON FORMATION

Field and laboratory studies reveal a close genetic relationship between Laurel gneiss and the Wissahickon formation. In order to demonstrate this relationship it will be necessary first to describe the latter briefly.

The Wissahickon formation which bounds the Laurel gneiss on the west is an oligoclase-mica schist. It consists chiefly of strongly folded mica schist interbedded with thin layers of fine-grained micaceous quartzite. A strong schistosity is characteristic of the mica schist. The quartzite, however, is only faintly banded by biotite-rich layers. These structures are regional and trend slightly east of north and dip steeply west (Pl. 1).

The mica schist is a completely recrystallized, coarse- to medium-grained rock which consists chiefly of muscovite, biotite, quartz, and calcic oligoclase ($Ab_{70}An_{30}$). Its grain size varies from 1 millimeter to several millimeters. Mica-rich layers (½ millimeter thick) alternate with layers rich in quartz and oligoclase (½ millimeter to 2 millimeters thick). The micas are in large flakes which are essentially oriented in a plane, but which locally wrap around large garnet crystals or lenticular areas of quartz. Practically all the quartz grains show strain

FIGURE 1. LAUREL GNEISS WITH REMNANTS OF FINE WISSAHICKON QUARTZITE
Remnants parallel to crude foliation of gneiss. Note that remnants look like inclusions in an igneous rock. Outcrop in Patuxent River 350 feet south of locality 35 (Pl. 1).

FIGURE 2. FAULTED QUARTZITE REMNANT WITH FOLIATION OF LAUREL GNEISS WRAPPING ABOUT IT
Note how the fault plane near the hand lens has been sealed over by later flowage and recrystallization. Appearance of gneiss is similar to that of an igneous rock. Outcrop at locality 35 (Pl. 1).

LAUREL GNEISS WITH REMNANTS OF WISSAHICKON QUARTZITE

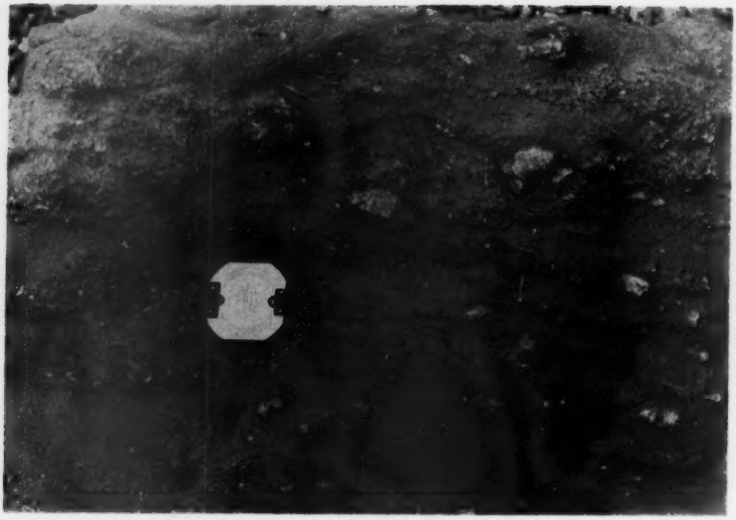


FIGURE 1. FINE-GRAINED LAUREL GNEISS WITH REMNANTS OF COARSE WISSAHICKON SCHIST
White, irregular quartz lumps are also remnants from Wissahickon formation. Picture suggests a hybrid rock produced by fine, even-grained granite penetrating coarse schist. Outcrop in Patuxent River 1000 feet northwest of locality 22 (Pl. 1).

FIGURE 2. LAUREL GNEISS WITH CONTORTED FLOW BANDS
Outcrop in Patuxent River 1000 feet northwest of locality 22 (Pl. 1).

LAUREL GNEISS WITH WISSAHICKON SCHIST REMNANTS, QUARTZ LUMPS, AND FLOW BANDS

shadows. Large porphyroblasts of muscovite, biotite, garnet, and tourmaline are common and show random orientations. They are filled with abundant inclusions of finer minerals of the groundmass. Accessory minerals of the schist include zircon, apatite, magnetite, and pyrite. Staurolite and cyanite have been found in this phase of the Wissahickon formation in some places, but they have not been observed close to the Laurel gneiss. Modes of the schist are shown in Table 1. It will be noted that the mineral percentages vary somewhat, although the same kinds of minerals are present in all specimens.

The quartzite phase of the Wissahickon formation is a light-gray, equigranular, faintly banded rock. Its grain size varies from ⅛ to ½ millimeter. The rock consists essentially of quartz and oligoclase with suboriented plates of muscovite and olive-brown biotite. Porphyroblasts of muscovite, biotite, and garnet are present in some specimens. Accessory minerals include zircon, apatite, magnetite, garnet, epidote, and chlorite. Modes of the quartzite are shown in Table 1.

RELATION OF THE LAUREL GNEISS TO THE WISSAHICKON FORMATION

MINERAL COMPOSITION

Table 1 shows the modes of three specimens of Laurel gneiss and several specimens of schist and quartzite from the Wissahickon formation near by. Five points regarding these should be emphasized.

(1) The three specimens of gneiss (5, 9, and 12A) show essentially the same minerals although in different proportions. It will be noted that the mineral assemblage in these specimens is not that of any typical igneous rock.

(2) The modes of the Wissahickon formation are variable in mineral proportions, but this is to be expected since they represent a sedimentary series which consists of alternating layers of schist and quartzite. Quartzite layers are not numerous in the Wissahickon formation, however, and are always thin. Detailed studies indicate that in the vicinity of Laurel, Maryland, the Wissahickon formation consists of approximately 90 per cent schist and 10 per cent quartzite. Specimens 1 and 11 were chosen because they are believed to be most representative of the schist phase of this formation. Specimens 20B and 25 were chosen as typical samples of the quartzite phase. The two schist specimens show a close modal similarity, but the two quartzite specimens differ somewhat in their mineral proportions. The quartzite specimens are richer in quartz and poorer in mica than the schist.

(3) There is an unquestionable similarity in certain respects between the modes of the Laurel gneiss (5, 9, and 12A) and those of the Wissa-

hickon formation. Exactly the same minerals occur in the gneiss that are present in the schist and quartzite. Furthermore, in the two samples of typical schist (1 and 11) even the relative proportions of these minerals are roughly the same as in the modes of the gneiss. However, the

TABLE 1.—*Modes of Laurel gneiss, schist, and quartzite from Wissahickon formation, and Guilford granite**

(Volume per cent)

	5	9	12A	1	11	16	20B	25	8
Quartz	40.9	48.2	37.7	43.8	40.9	45.2	66.6	65.0	30.2
Plagioclase	16.3	22.2	21.5	19.2	20.1	13.1	9.4	16.5	33.5
	An_{32}	An_{28}	An_{30}	An_{28}	An_{28}	An_{27}	An_{28}	An_{28}	An_{10-15}
Biotite	16.6	11.2	11.5	12.3	14.0	15.2	4.8	8.4	3.5
Muscovite	24.1	16.9	28.2	22.4	23.0	22.4	15.7	7.8	tr
Garnet	0.7	0.7	....	1.6	1.5	3.7	3.3	....	....
Magnetite	1.4	1.0	1.1	0.7	0.5	0.4	0.2	1.8	tr
Tourmaline	....	....	....	tr	....	tr	....	....	....
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	tr	....	tr	tr
Chlorite	tr	tr	tr	tr	tr	tr	tr	....	....
Epidote	....	....	....	....	....	....	....	0.5	....
Microcline	....	....	....	....	....	....	....	....	32.8

* All modes measured with a Wentworth integrating stage.

All locality references are to Plate 1.

5 Laurel gneiss, 800 feet west-southwest of locality 1.
9 Laurel gneiss, locality 9.
12A Laurel gneiss, locality 20.
1 Schist from Wissahickon formation, locality 1.
11 Schist from Wissahickon formation, 1/3 mile west-southwest of locality 9.
16 Schist-gneiss, locality 28. Transition product between Laurel gneiss and schist of Wissahickon formation.
20B Quartzite from Wissahickon formation, locality 33.
25 Quartzite from Wissahickon formation, near locality 33.
8 Guilford granite, locality 8.

quartzite specimens (20B and 25) differ considerably from the gneiss in mineral ratios. These relations are exactly as would be expected if, as the writer believes, the Laurel gneiss has been derived from the Wissahickon formation. By granulation, flowage, and recrystallization the heterogeneous Wissahickon formation has been thoroughly mixed to produce a more or less homogeneous rock. Since the quartzite layers are not numerous (about 10 per cent of the Wissahickon formation), they would not make the gneiss much richer in silica than the typical schist.

(4) Specimen 16 is from the transition zone between the Laurel gneiss and the Wissahickon formation. It is believed to represent schist partly transformed to gneiss. Its mode is similar to those of both these rock types.

FIGURE 1. FINE-GRAINED, GRAY, FOLIATED LAUREL GNEISS
Enclosing elliptical areas which consist of mica segregations surrounded by cream-colored gneiss. Outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

FIGURE 2. FINE-GRAINED, GRAY LAUREL GNEISS
With areas of light-colored gneiss containing mica segregations. Segregations have formed at the expense of the micas in the normal gray gneiss, thus producing light areas. Outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

MASSES OF SEGREGATIONS IN THE LAUREL GNEISS

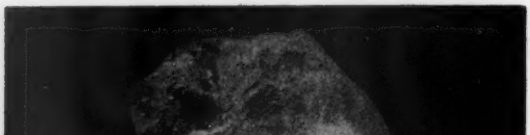


FIGURE 1. SPECIMEN OF LAUREL GNEISS (gray)
Containing mica segregations (black spots) surrounded by cream-colored zone. Segregations consist of biotite and muscovite porphyroblasts accumulated by aqueous solutions. Material of the segregations was derived from groundmass micas of gneiss, thus leaving a light-colored zone around each segregation. Height of specimen is 3½ inches. From outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

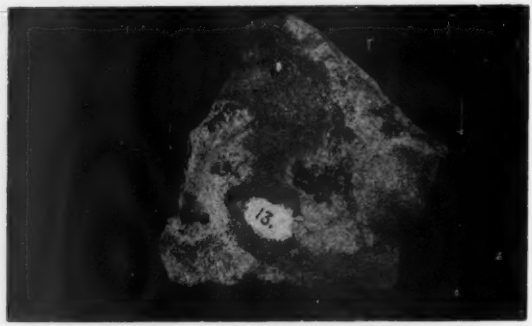


FIGURE 2. SAME AS FIGURE 1 OF PLATE 5
Height of specimen is 4 inches.

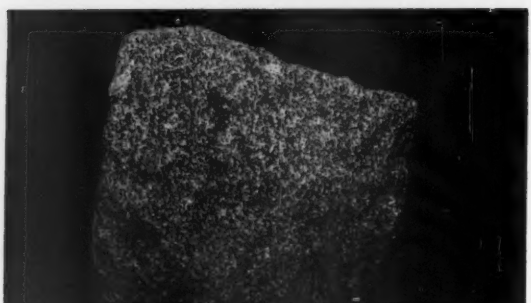


FIGURE 3. SPECIMEN OF TYPICAL LAUREL GNEISS
Note apparent similarity to a plutonic igneous rock. Height of specimen is 3½ inches. From locality 20 (Pl. 1).

MICA SEGREGATIONS AND TYPICAL LAUREL GNEISS

(5) The optical properties of the plagioclase, muscovite, biotite, garnet, and tourmaline in the Wissahickon formation have been determined accurately. These optical properties are identical with those of the corresponding minerals in the gneiss. This indicates that the minerals of the two rocks are similar in chemical composition.

Spectrographic determinations were made of the alkali content of the biotite from both the Laurel gneiss and the schist from the Wissahickon formation. Biotite was separated from specimen 12A of the gneiss from locality 20, and from a specimen of coarse schist from an outcrop 7 miles northwest of the northern end of the gneiss body (1 mile northwest of the village of Clarksville). The alkali content was found to be the same in both biotites within the range of spectrographic methods. The results are as follows:

Material	Approximate percentage
K_2O	5
Rb_2O	1
Li_2O	1
Na_2O	3
Cs_2O	not detected

This evidence corroborates the optical data in indicating that the biotite in the two rocks is the same.

CHEMICAL COMPOSITION

In comparing the chemical composition of the Laurel gneiss with that of the Wissahickon formation it was found impracticable to make chemical analyses of the rocks. From rocks which are as variable as these two, it would be impossible to select a few specimens for analyses which would be thoroughly representative of each type. Many chemical analyses of well-chosen specimens would be necessary in order to determine definitely any chemical relationship between the Laurel gneiss and the Wissahickon formation. Such chemical analyses would involve an expense all out of proportion to the benefits derived.

The author has thought of using composite samples, each consisting of approximately 20 pieces of equal weight, of the two principal rock types: Laurel gneiss and Wissahickon schist. However, this possibility was discussed with a number of leading petrographers, and they all recognized the problem of proper sampling. Plates 2, 3, and 4 will readily convince anyone of the difficulties of sampling in rocks of which no 2 cubic feet are alike. Inasmuch as all the minerals in both rocks are exactly of the same composition and there are none in one rock that do not occur in the other, chemical analyses could at best show the differences in proportion of the otherwise identical minerals. There is little

doubt in the author's mind that analyses would show nothing but would merely indicate the chemical similarity of these two rock types.

For these reasons it was considered more practicable to calculate the average chemical composition of the Wissahickon formation and the

TABLE 2.—*Chemical compositions*

	A	B	C
SiO_2	72.87	71.10	70.18
TiO_2	.....	.....	0.39
Al_2O_3	16.27	17.10	14.47
Fe_2O_3	0.48	0.76	1.57
FeO	1.54	1.72	1.78
MnO	.....	.....	0.12
MgO	1.08	1.13	0.88
CaO	1.07	1.21	1.99
Na_2O	1.61	1.65	3.48
K_2O	3.95	4.29	4.11
P_2O_5	.....	.....	0.19
H_2O	1.13	1.04	0.84
	100.00	100.00	100.00

A Average chemical composition of the Wissahickon formation near Laurel, Maryland. Calculated from modes of schist and quartzite shown in Table 1.

B Average chemical composition of Laurel gneiss. Calculated from the three modes shown in Table 1.

C Chemical composition of average granite of all periods (Daly, 1933, p. 9).

Laurel gneiss from a study of several thin sections. Since the mineralogy of both rock types is relatively simple, the results obtained are believed to be fairly accurate.

After a study of several thin sections, specimens 1 and 11 (Table 1) were chosen as most representative of the schist phase of the Wissahickon formation, and specimens 20B and 25 were chosen as representative of the quartzite phase. It was believed better to choose representative thin sections rather than to average all of them since some sections were prepared from unusual rock types. The modes of specimens 1 and 11 were then averaged to give the average mode of the schist. Likewise, the average mode of quartzite was obtained by averaging the modes of specimens 20B and 25. Now it has been pointed out that the schist makes up about 90 per cent of the Wissahickon formation, and the quartzite about 10 per cent. Thus by taking the average mode of the schist and the average mode of the quartzite, in a ratio of 9 to 1, the average mode of the whole Wissahickon formation was obtained. From this average mode, the average chemical composition of the Wissahickon formation was calculated (Table 2, column A).

The average chemical composition of the Laurel gneiss was determined in a similar way. Several thin sections of the rock were studied, and specimens 5, 9, and 12A (Table 1) were chosen as representative of the whole body. The modes of these three specimens were then averaged so that the average mode of the gneiss was obtained. From this mode, the average chemical composition of the Laurel gneiss was calculated (Table 2, column B).

A study of Table 2 shows a surprisingly close similarity between the calculated average chemical composition of the Laurel gneiss (column B) and that of the Wissahickon formation (column A). The author's contention that the Laurel gneiss was produced by the thorough mixing and recrystallization of the various rock types of the Wissahickon formation is fully in line with this evidence.

Column C in Table 2 gives the chemical composition of the average granite of all periods as calculated by Daly. It will be noted that the composition of the Laurel gneiss (column B) is similar to that of the average granite except that alumina is somewhat higher and soda somewhat lower in the former. The significance of this point will be noted later.

STRUCTURAL AND AGE RELATIONS

The foliation of the Laurel gneiss is essentially parallel to that of the Wissahickon formation (Pl. 1). The following observations indicate, however, that the gneiss and its foliation are younger than the Wissahickon formation and its foliation. In the first place, the schist of the Wissahickon formation is a coarse, thoroughly recrystallized, mica-rich rock with a well-developed schistosity or foliation. This foliation is secondary and has been produced by intense regional metamorphism. The Laurel gneiss, on the other hand, does not show this marked schistosity so it must be younger. It is finer-grained and shows only a vague foliation which has more the appearance of a primary flow structure. In the second place, remnants of quartzite and strongly foliated schist from the Wissahickon formation are common in the gneiss. Many of these are oriented so that their foliation is parallel to the poorer foliation of the gneiss; others are unoriented. Finally, at localities 11 and 20 (Pl. 1) the foliation of the gneiss bends around remnants of quartzite which are themselves banded. (*See* discussion of quartzite remnants.) The foliation of the Laurel gneiss has clearly been deformed around the older banded remnants.

FABRIC ANALYSES

By Ernst Cloos

Introduction.—Fabric studies of the Laurel gneiss and Wissahickon schist indicate three significant things. First, the mica orientation in

the schist and in the gneiss are essentially the same. This suggests that the directions of movement which produced mica orientation in the gneiss are nearly the same as those which caused orientation in the schist. Secondly, the c axes of quartz in the schist are essentially perpendicular to the foliation of the schist. In the gneiss, however, the largest concentration of quartz axes lies in the foliation planes produced by the micas, whereas a second group is oriented at an angle of about 45 degrees to these planes. It appears that the quartz in the gneiss represents original large grains which have been fractured along crystal directions and oriented according to shape by rock flowage. This supports the concept that the Laurel gneiss represents the Wissahickon formation after granulation, plastic deformation, and later recrystallization. Finally, the fabric of the schist is definitely that of a metamorphic rock (B-tectonite), whereas the quartz orientation in the Laurel gneiss is similar to that in an igneous rock.

Method.—Biotite, muscovite, and quartz orientations were measured with the universal stage in the conventional manner. The lower half of an equal-area projection was used, and counting at each half-centimeter intersection was found more accurate than the usual method of counting at each centimeter intersection. The preparation and interpretation of fabric diagrams is adequately described in the literature by Haff (1938), Knopf and Ingerson (1938), and Fairbairn (1939), and a detailed account of the method seems superfluous here.

One specimen each of schist from the Wissahickon formation (locality 33, Pl. 1) and the Laurel gneiss (450 feet northeast of locality 33, Pl. 1) were investigated. Thin sections were cut as closely as possible at an angle of 90 degrees to the strike of the megascopic foliation. Strike and dip of the projection planes are marked with an arrow and a short line in the diagram. East and west are marked E and W respectively. All strike readings were measured clockwise from N (= 0°) to S (= 180°). For example: N 45°E = 45°, and N 45°W = 135°, with the arrow pointing northeast and northwest respectively.

The mica diagrams include 100 grains each, and since the orientation is very pronounced this number is thought to be sufficient in ascertaining the principal orientation of cleavage poles. In the schist 400 average-sized quartz grains were measured, and 200 large quartz grains were measured in the gneiss.

Figure 5 shows the orientation of 100 quartz axes in relation to the shape of the quartz grains in the gneiss. Since many of the grains are elongated, the dimensions were measured with a micrometer, and the

ratio between the longest and shortest form axes was plotted in the location of the optical axes. In this manner the shape of each grain appears in relation to its optical orientation. The diagram is not a counted diagram.

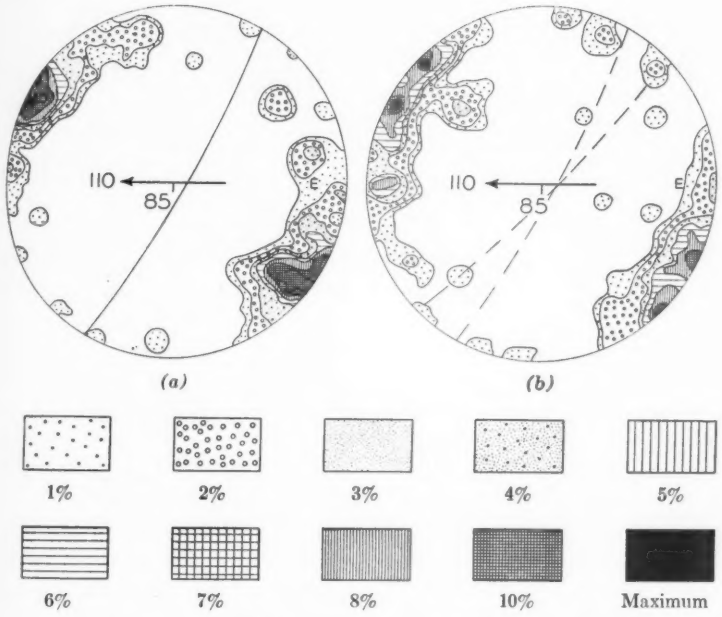


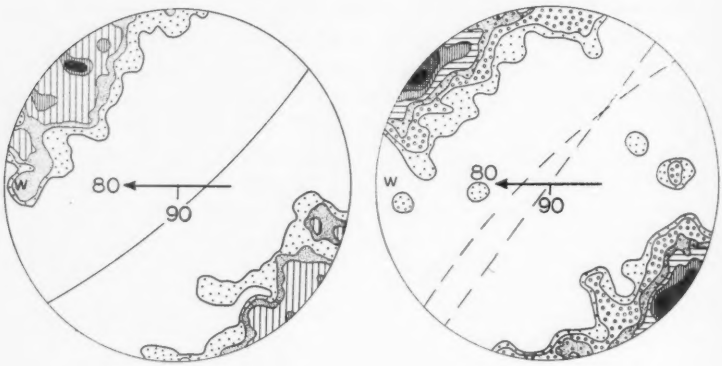
FIGURE 2.—*Muscovite and biotite diagrams of Wissahickon schist*

(*a*) Muscovite from Locality 33 (Pl. 1). 100 muscovite cleavage poles, excluding porphyroblasts. Contours 1-2-4-6-8-10-12(15). Trace of cleavage plane indicated by solid line.

(*b*) Biotite from Locality 33. 100 biotite cleavage poles. Contours 1-2-4-6-8-10-12(13). Traces of biotite cleavages indicated by broken lines. Girdle interrupted, axis of rotation about 10° E. of center.

Mica orientation.—All micas measured are those in the groundmass; no porphyroblasts were included. The poles of biotite cleavages in both rocks are concentrated in two distinct maxima, representing two planes which intersect at angles of 10 and 15 degrees (Figs. 2b, 3b). A girdle is suggested but is incomplete. In spite of the small angle between the planes, and the high concentration around the maxima, the writer believes that the concentration of biotite in two planes is significant since this freedom of orientation has been found also in more than 100 other localities throughout the Wissahickon formation.

The muscovite cleavage poles in both rocks are concentrated in only one maximum (Figs. 2a, 3a) which does not necessarily coincide with any of the biotite maxima. In other areas of the Wissahickon formation large deviations up to 30 degrees have been measured, and these are



(a) *(b)*

FIGURE 3.—*Muscovite and biotite diagrams of Laurel gneiss*

(*a*) Muscovite from 450 feet northeast of locality 33 (Pl. 1). 100 muscovite cleavage poles. Contours 1-3-5-8-10. Trace of cleavage indicated by solid line.

(*b*) Biotite from same locality as 3a. 100 biotite cleavage poles. Contours 1-2-4-6-8-10(12-16). Broken lines indicate traces of biotite cleavages intersecting in quartz maxima of Figure 4b.

See Figure 2 for contour shading.

thought to be due to a rotation of the fabric between stages of recrystallization. In these areas muscovite is later than biotite, and its orientation is always more distinct (Cloos and Hietanen, 1941).

In the present investigation no age difference between biotite and muscovite was noted, and this may explain the unusually small deviation between orientations. As a whole, biotite is not so well oriented as muscovite. The arrangement of mica poles in both schist and gneiss represents a well-defined megascopic schistosity and foliation respectively (Figs. 2a-3b).

Quartz orientation.—The orientation of quartz grains differs widely in the two rock types (Figs. 4a, 4b). In the schist (Fig. 4a) 400 average-sized grains are concentrated in an incomplete girdle which covers half the circumference and culminates in a 4 per cent maximum, near the location of the cleavage poles of the micas. This shows preferred orientation of quartz grains roughly perpendicular to the cleavage planes. This orientation is described by Fairbairn (1939) as maximum V and

is believed to be due to yielding of the grains parallel to the base (0001) during deformation. In other areas where the Wissahickon formation has been studied a similar arrangement is common and indicates probably the typical quartz orientation in a tectonite in reference to an

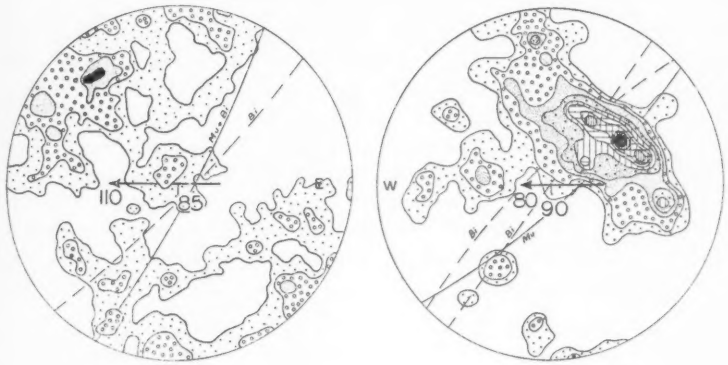


FIGURE 4.—*Quartz diagrams of Wissahickon schist and Laurel gneiss*

(*a*) Wissahickon schist. Locality same as 2a. 400 average size quartz grains. Contours 1-2-3-4. Cleavage traces from Figures 2a and 2b. Mu = muscovite; Bi = Biotite.

(*b*) Laurel gneiss. Locality same as 3a. 200 large quartz grains. Contours 1-2-3-4-5-6-7-8. Main maxima in muscovite cleavage (Mu) and the intersection of biotite cleavages (Bi).

See Figure 2 for contour shading.

axis of rotation. The axis emerges about 10 degrees east of the center of the diagram and at the intersection of muscovite and biotite cleavage planes. The incompleteness of the quartz girdle is in harmony with the lack of a mica girdle.

Figure 4a was prepared in two stages of 200 grains each, and the maxima of the two partial diagrams reached 8 per cent. The total diagram is otherwise identical with each of the two partial ones. The present orientation shown in Figure 4a was thus fixed after 200 grains had been plotted.

In the Laurel gneiss (Fig. 4b) quartz axes are concentrated in a steep 8 per cent maximum which is in the muscovite cleavage plane. Since large and elongated quartz grains were measured, the author assumes that this orientation is due, in part at least, to the shape of the grains and not entirely to deformation and recrystallization. In order to avoid a concentration of axes of portions of one large, broken quartz grain in a false maximum, individual grains which were close together were

counted as one. In addition, the dimensions of grains were measured, and the ratios plotted in Figure 5.

It is obvious that the concentration of points in Figure 5 is identical with the maxima in diagram Figure 4b. A close relation between grain

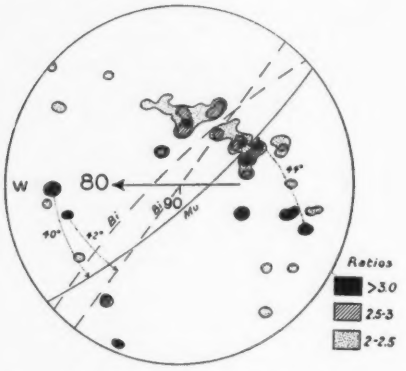


FIGURE 5.—*Quartz diagram of Laurel gneiss*

Locality same as 3a. Ratio of longest over shortest dimension of 100 quartz grains plotted in the axes points of Figure 4b. Solid and broken lines represent cleavage traces. Bi and Mu indicate traces of biotite and muscovite cleavage planes respectively.

shape and orientation thus becomes evident since those grains with the highest ratios (3:1 and above) are located in the principal maxima.

Interpretation of quartz orientation.—All except two of the high ratio areas in Figure 5 coincide with important quartz maxima. The largest concentration is in the cleavage planes of muscovite (Mu) and biotite (Bi). These quartz grains are thus elongated parallel to the prism (1010) which in turn is very near or within the foliation plane. A second group of grains is concentrated in areas which are 40 to 45 degrees from the foliation planes (40, 42, and 44 degrees in Fig. 5).

Since all these quartz grains are broken into many elongated fragments, the fracturing seems to have taken place largely along "cleavages" which can be identified as mineralogical directions: base, prism, and rhombohedron. Their later orientation in the Laurel gneiss took place predominantly according to shape. The large quartz grains indicate flow structures, and their orientation appears to represent flow lines. Similar orientations are well known from many igneous masses in which fragments or elongated crystals are orientated parallel to flow lines or

planes (Balk, 1937). The author believes that the fragmentation of quartz took place during an early stage of granulation and that the fragments were oriented later largely according to form. Fragmentation took place along quartz cleavages and parallel to common crystal directions.

TEXTURAL RELATIONS

The texture of the Laurel gneiss is quite different from that of the schist from the Wissahickon formation. The schist is a coarse, foliated rock which consists of bands of quartz and oligoclase alternating with layers of coarse, well-oriented micas. The gneiss, on the other hand, is a fine-grained, sugary rock with a crude alignment of biotite and muscovite. The micas are scattered throughout the rock and are not localized in bands as in the schist. The texture of the schist, therefore, is that of a strongly metamorphosed rock, whereas the megascopic texture of the gneiss appears more like that of an igneous type.

GRADATIONAL CONTACTS

Wherever exposed, the boundary between the Laurel gneiss and the Wissahickon formation is gradational; in many places it is difficult to establish. In the vicinity of locality 1 (Pl. 1), for example, the Wissahickon schist grades into typical Laurel gneiss within a distance of about 300 feet. A similar relation is shown in a traverse across the contact from locality 28 to locality 33 along the Patuxent River (Pl. 1). Here the rock changes gradually from a fine, granular, poorly foliated type (typical Laurel gneiss) into the coarse, foliated mica schist and interbedded quartzite of the Wissahickon formation. The transitional relation may also be studied at locality 56 (Pl. 1).

REMNANTS

Another important relationship is shown by the presence of remnants of schist and quartzite in the Laurel gneiss. The schist remnants show irregular boundaries across which there is a gradual transition from schist into gneiss. Even under the microscope it is difficult to determine where one ends and the other begins. The change is entirely textural since the mode of the schist remnants and that of the gneiss are essentially the same.

ORIGIN OF THE LAUREL GNEISS

GENERAL STATEMENT

Only two explanations of the origin of the Laurel gneiss appear plausible. It may be the result of "granitization" of the Wissahickon formation by an intrusive magma or it may have originated by metamorphism of the Wissahickon formation.

HYPOTHESIS OF GRANITIZATION

Cloos interpreted the Laurel gneiss as a migmatite—the result of granitization of the Wissahickon formation. He believed that the latter was actually penetrated, dissolved, and assimilated by intrusions of granitic magma. Cloos' hypothesis was based upon field observations, and the Laurel gneiss, as mentioned above, has many of the megascopic features of an igneous rock. It might also be assumed that the Wissahickon formation was granitized through the introduction of material by solutions from a granitic magma rather than by the magma itself.

Some arguments in favor of granitization of the Wissahickon formation are as follows: (1) The normal Laurel gneiss has the appearance of a granular igneous rock with flow structure; (2) it has a chemical composition somewhat similar to that of a true granite (Table 2); (3) it contains abundant fragments of schist and quartzite from the Wissahickon formation, in some places so many that the gneiss looks like a hybrid rock; and (4) a granitic intrusive (Guilford granite) bounds the Laurel gneiss on the north.

A number of facts, however, individually suggest that the Laurel gneiss is not the result of granitization. None of these arguments is, in itself, a definite proof, but when considered together they indicate that the gneiss was not formed by granitization.

(1) Sharp intrusive boundaries between the Laurel gneiss and the Wissahickon formation have never been found. This means that the former could not have been formed by the injection of granitic magma into the Wissahickon formation, otherwise one would expect to find at least a few dikes and sills penetrating the schist. On the other hand, this argument does not preclude the possibility of granitization by solutions since this process would normally produce gradational boundaries.

(2) The Laurel gneiss is, as a rule, finer-grained than the Wissahickon formation. If the Wissahickon formation had been granitized, one would expect the resulting rock to be at least as coarse as, if not considerably coarser than, the original rock. Recrystallization is an important feature of granitization, and this increases grain size.

(3) No mineral occurs in the gneiss that has not been found in the Wissahickon formation in the surrounding area. Furthermore, those minerals in the former are exactly the same optically (and, therefore, chemically) as those in the schist and quartzite. It would be a strange coincidence indeed if the minerals of an igneous intrusive, or those introduced by solutions, were exactly the same as those of the country rock. It is admitted that this might be possible if considered in the light of the principle of mineral facies. Thus an igneous rock and a meta-

morphic rock may contain the same mineral assemblage providing that both had the same original chemical composition and that both formed under similar temperature-pressure conditions. However, it seems very unlikely that the intrusion would have crystallized under exactly the same temperature-pressure conditions as those under which the Wissahickon formation was metamorphosed. It appears, therefore, that the similarity in mineral composition between the Laurel gneiss and the Wissahickon formation is significant evidence against the hypothesis of granitization.

(4) The so-called Guilford "granite" (really a granodiorite), which is the only exposed rock close enough to have caused the granitization, could not have had anything to do with the origin of the Laurel gneiss for several reasons. In the first place, the Guilford granite is younger than the gneiss. Numerous dikes of it cut the gneiss near locality 1 (Pl. 1), and a wide dike occurs in the stream at locality 60 (Pl. 1). Their boundaries are always sharp and distinct and are never gradational or characterized by stringers and inclusions. Coarse, pink pegmatites from the Guilford granite cut the gneiss along clean, sharp boundaries in several localities. Secondly, the Guilford granite contains a high percentage of potash feldspar. A mode of this rock (8) is shown in Table 1. In all the sections of gneiss examined, however, no potash feldspar has been found. In the third place, the Guilford granite is texturally unlike the gneiss. It is a light-gray, nonfoliated, biotite rock with a grain size of from 1 to 2 millimeters. The gneiss, it will be remembered, is faintly foliated and finer-grained.

Because of these facts, and because the Laurel gneiss can be explained more satisfactorily in another way, the writer does not believe that the hypothesis of granitization of the Wissahickon formation either by an intrusive rock or by solutions can be accepted.

HYPOTHESIS OF METAMORPHISM

Introduction.—A completely satisfactory explanation of the origin of the Laurel gneiss must take into account the following pertinent facts: (1) The mineralogical and chemical compositions of the Laurel gneiss are essentially the same as those of the Wissahickon formation; (2) wherever the contact can be studied there is a complete textural gradation from the one rock type into the other; (3) the gneiss is younger than the Wissahickon formation and contains abundant fragments (remnants) of schist and quartzite; (4) the gneiss has undergone both granulation and recrystallization, and the structure and fabric indicate that it has flowed as a plastic mass. After considering these facts, the writer is convinced that the Laurel gneiss originated by the granulation, flowage,

and recrystallization of the Wissahickon formation. Granulation took place under shearing stress in the early stage of the process when the rocks were cold and dry; flowage and recrystallization occurred later under conditions of higher temperature and abundant water. It is be-

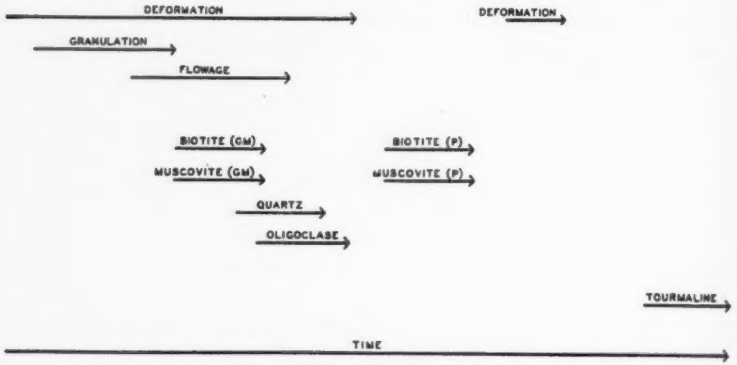


FIGURE 6.—*Graph showing time relationships between the various stages in formation of the Laurel gneiss*

GM = groundmass, P = porphyroblasts.

lieved that the complete transformation took place without appreciable addition or loss of material.

The various stages in this process will be described below in detail. The time relationships between the different stages is shown graphically in Figure 6.

Granulation.—The evidence for granulation is now largely obliterated by recrystallization. That this was once an important process in the formation of the Laurel gneiss, however, is shown in several ways:

(1) The typical Laurel gneiss is finer-grained than the schist of the Wissahickon formation. Since recrystallization ordinarily tends to increase grain size, it would suggest that the schist was crushed and granulated before it was recrystallized.

(2) The strongest and most direct evidence of granulation is furnished by the universal straining and breaking of quartz grains in the gneiss. (*See* petrographic description and discussion of fabric.) Many large crystals of quartz have been strained almost to the breaking point, whereas others have actually broken into smaller fragments along irregular cracks. Many of these fragments have been displaced relative to the large crystal, but some are still sufficiently close in optical parallelism so that there can be no doubt that they represent parts of a once complete crystal.

In the interior of the gneiss body the granulation of the quartz is more or less complete, but in the transition zone between the Laurel gneiss and the Wissahickon formation there are numerous large grains of quartz which have not been fully broken down (Pl. 6, fig. 4).

(3) At least one quartzite remnant has been faulted (Pl. 2, fig. 2), and the fault plane healed by later flow and recrystallization.

The writer believes that the granulation took place during the early stages of transformation by the complete mechanical destruction ("Durchbewegung") of the Wissahickon formation. In these early stages the rocks were cold and dry and yielded by granulation and crushing. The large quartz grains in the schist were strained and broken into smaller grains. The oligoclase crystals were probably also crushed and granulated in a similar manner although little evidence of this can be seen in the gneiss today. The micas were bent and twisted, and both they and the garnets were broken into smaller fragments. Thus, by this process the coarse Wissahickon formation was transformed into a finer cataclastic rock. Locally, where stress conditions were less severe, the schist and quartzite were not completely granulated, and remnants of these are now preserved in the gneiss.

Flowage.—With the increase of temperature and solutions, the rock mass began to yield by flowage as well as by granulation. Individual grains of quartz, oligoclase, and mica moved about relative to one another, partly by slipping along crystal boundaries and partly by plastic deformation of grains. Mica crystals became crudely oriented in response to differential movements in the mass. Many remnants of schist and quartzite which were not previously granulated were twisted about into random orientations. Continued granulation accompanied this process, but, as the temperature rose and water increased in amount, flowage became the dominant mode of adjustment to the stress, and crushing and granulation became less important.

Rock flowage did not cease at the end of the period of crushing and granulation but continued on well into the period of recrystallization. This is certain because many recrystallization features show evidences of this flowage: (1) Mica segregations are pulled out into streaks and bands; (2) the foliation, which is caused by roughly oriented recrystallized micas, is locally thrown into complex folds; (3) the foliation wraps around many quartzite remnants in a manner suggesting that it actually flowed about them; and (4) the fabric of the gneiss, which is at least partly a result of recrystallization, shows unquestionable evidence of flowage. The late stages of rock flowage and the early stages of recrystallization were thus contemporaneous.

Recrystallization.—After granulation had largely ceased, but while some flowage was still in progress, recrystallization of the crushed rock began. Although stress still operated, it was becoming relatively less important, and heat and water were becoming the dominant factors. Large grains grew at the expense of small ones, strained crystals tended to dissolve and unstrained ones to take their place, and micas grew in directions of least resistance or were oriented by flowage so that the rock took on a faint foliation. Thus the rock lost much of its crushed and granulated appearance.

Recrystallization of the essential minerals may be divided into two stages: (1) formation of minerals of the groundmass, and (2) formation of porphyroblasts.

In the first stage quartz, plagioclase, and the oriented micas formed. It is clear that the quartz and plagioclase are essentially contemporaneous in their crystallization since both minerals are mutually interlocked by sutured boundaries. The oriented micas formed either slightly before or during the recrystallization of the quartz and plagioclase. Their orientation is due probably to two factors—*i.e.*, crystallization under stress and rock flowage.

The porphyroblasts formed during a later stage of recrystallization, after the stress had ceased entirely. The muscovite and biotite porphyroblasts cut across the foliation of the gneiss and inclose poikiloblastically crystals of apatite, magnetite, zircon, quartz, plagioclase, and small crystals of mica. Figure 2 of Plate 6 shows a muscovite porphyroblast with inclusions of poorly oriented biotite. It has already been pointed out that the muscovite and biotite porphyroblasts are mutually intergrown in many cases, suggesting simultaneous crystallization. They show undulatory extinction due to straining and prove that a period of renewed stress followed their formation. These events were followed by the crystallization of scattered porphyroblasts of tourmaline, some of which appear to cut directly across mica porphyroblasts. The position of the garnets in the sequence is not entirely clear. Most of them appear to be remnants of coarse porphyroblasts which formed originally in the schist and were not completely broken down by granulation. A few of them have undergone some recrystallization.

It has been shown that the musovite and biotite porphyroblasts have the same optical properties as the fine, oriented muscovite and biotite in the groundmass, and that their chemical compositions are, therefore, probably the same. It has also been pointed out that the porphyroblasts originated later than the groundmass micas through the medium of hot aqueous solutions. There is no evidence whatever that any material was added to or removed from the gneiss by these solutions; they appear

to have acted simply as agents for recrystallization by dissolving the groundmass micas and depositing them as porphyroblasts. Barth (1936, p. 805-806) has advocated a somewhat similar explanation for the muscovite metacrysts in the porphyroblastic muscovite schist of Dutchess County, New York.

Although granulation, flowage, and the early stages of recrystallization were confined to the area of the Laurel gneiss, the formation of late porphyroblasts extended out into the Wissahickon formation. This suggests that in the later stages of recrystallization solutions had become effective over a large area.

The segregations of mica porphyroblasts with their surrounding areas of cream-colored rock also belong to this late stage of recrystallization. It is thought that they originated in essentially the same way as the scattered porphyroblasts. In this case, however, the process was locally more complete. Practically all groundmass micas were dissolved from small areas of the gneiss, thus leaving the rock cream-colored. The dissolved material was then reprecipitated as porphyroblasts around specific centers.

Agencies of metamorphism.—The agencies of metamorphism were: (1) shearing stress, (2) heat, and (3) solutions (water).

The stress causing granulation and flowage is believed to represent a late phase of the regional stress which produced the foliation in the Wissahickon formation. The general conformity of the foliation in the Laurel gneiss with that in the Wissahickon formation suggests that the late stress phase acted in essentially the same direction as the main stress. We have no way of knowing the exact shape of the gneiss body since much of it is now covered by Cretaceous sediments. However, in general, it is long and narrow and trends northeast-southwest parallel to the regional strike. This suggests that the gneiss developed along a broad fault zone or shear zone where crushing and granulation were most intense.

The heat necessary for flowage and recrystallization is believed to have originated in three ways: (1) heat from the geothermal gradient, (2) heat generated mechanically by crushing and granulation, and (3) heat brought up by conduction and by solutions from a magma below.

The temperature at which the gneiss formed can be determined in a general way if two assumptions are made: (1) The Wissahickon formation was overlain by all the Paleozoic formations at the time of its metamorphism (Permian?), and (2) the gneiss formed shortly after or during the latest phase of the metamorphism of the Wissahickon formation before the load was appreciably reduced. Ernst Cloos (personal

communication) estimates that the Paleozoic column above the gneiss must have had a minimum thickness of 30,000 feet or 9 kilometers. According to Adams (1924, p. 468, Fig. 2), the temperature at this depth, resulting from the geothermal gradient, is approximately 250° C. If we add to this the heat generated by crushing and granulation, and that introduced by solutions and by conduction from a magma below, we might at first conclude that the temperature was several hundred degrees. However, there is nothing to indicate that the temperature was exceptionally high. For example, no portion of the rock shows any evidence of fusion. Furthermore, recrystallization was not complete since much of the quartz is still strained. More specifically, the absence of orthoclase suggests that the temperature during recrystallization was not above 340° C. (Grout, 1932, p. 441). Roughly, the temperature during metamorphism probably ranged between 250° and 340°.

The writer believes that water played an important role in the flowage and recrystallization of the Laurel gneiss. This is suggested in two ways: (1) Contorted foliation indicates that the gneiss was, at least locally, a very plastic mass. The importance of water in increasing the plasticity of rocks is well recognized. (2) The abundance of porphyroblasts and the occurrence of irregular segregations of mica suggest that water penetrated the gneiss thoroughly during the final stage of transformation. (*See* section on recrystallization.) The water is thought to have had two sources: (1) water liberated from the micas during the metamorphism, and (2) water rising from below along a fault zone or shear zone. The presence of tourmaline as late porphyroblasts in both the gneiss and the schist suggests an igneous body at depth from which both the water and heat emanated. Except for scattered crystals of tourmaline, there is no evidence whatever that material was added or subtracted during recrystallization. Both the Laurel gneiss and the Wissahickon formation have essentially the same chemical compositions. This suggests that the water promoting recrystallization was relatively pure.

PETROGENETIC IMPORTANCE OF THE LAUREL GNEISS

The problem of the Laurel gneiss raises a question of fundamental importance in the science of petrogenesis. Here is a rock whose general characteristics and relation to the surrounding rocks indicate that it was formed simply by the metamorphism of the Wissahickon formation without the addition or subtraction of appreciable material. And yet this rock possesses most of the features of a typical igneous rock. It has the structure and megascopic texture of an igneous rock, and a chemical composition similar to that of granite. It likewise contains segregations which are not unlike those found in many igneous bodies. Ernst Cloos

has shown that the fabric of the gneiss is somewhat similar to that of an igneous rock which has flowed. Furthermore, the gneiss contains fragments of schist and quartzite which might easily be interpreted as inclusions engulfed in a magma. In fact, where these remnants are abundant, the Laurel gneiss closely resembles a migmatite, and the writer, therefore, suggests that it be called a "pseudomigmatite."

There are two differences, however, between the Laurel gneiss and an igneous rock: (1) The mineral composition of the gneiss is not that of any typical igneous rock (but is the same as the wall rock), and (2) the gneiss shows gradational contacts with the adjacent Wissahickon formation.

The chief mineralogical difference between the Laurel gneiss and a typical granite is that muscovite occurs instead of orthoclase. However, if the temperature of metamorphism had been higher (probably above 340° C.), orthoclase might have formed instead of muscovite. Then the gneiss would have had not only the appearance of a granite and a somewhat similar chemical composition, but also the mineral composition of a true granite.

Under somewhat different conditions of metamorphism the Laurel gneiss might also have developed intrusive contacts. Any agent which could increase the plasticity or mobility of the rock might cause it to flow more readily and thereby intrude the surrounding rocks. Intense stress would, without doubt, be an important agent. Then too, Griggs (1940) has clearly demonstrated that water is of fundamental importance in rock flowage. The effect of high temperature is somewhat debatable, but both field data (Grout, 1932, p. 390) and some experimental data (Adams, 1912, p. 111-112) indicate that, in general, rocks yield more readily at high temperatures. Let us suppose that during transformation the stress had been more intense, the temperature had been higher, and solutions had been more abundant. Under these conditions the gneiss would have been more plastic and more mobile and might easily have penetrated cracks and crevices in the surrounding country rock to form dikes and sills with true igneous boundaries.

It appears, then, that the Laurel gneiss might have had all the features of a true igneous rock if the agencies of metamorphism had been more intense. Under these conditions it would have been impossible to distinguish it from an igneous rock by any criterion.

In recent years much has been written on the subject of igneous-looking rocks which have presumably been formed by some process other than the crystallization of magma (Goldschmidt, 1921; Sederholm, 1926; Eskola, 1932, 1933; Gilluly, 1933; Barth, 1936; Wegmann, 1936; Anderson, 1937; C. A. Chapman, 1939, p. 167-170; Postel, 1940). For

example, "syntexis" is the process by which crystalline or sedimentary rocks are either penetrated by magma or altered by magmatic solutions until, in some cases, they look like true igneous rocks. "Ultrametamorphism" is the supposed process by which a solid rock becomes so intensely metamorphosed that it is fused, either selectively or completely, and then crystallized again to produce an igneous-looking rock. Many variations of these processes have been described in the literature. The difficulties in distinguishing between igneous-looking rocks produced by such processes and true igneous rocks which have crystallized from magmas are now well recognized in petrogenesis. They have been briefly summarized in a recent paper by Grout (1941).

Like some of the syntetic rocks, the Laurel gneiss has most of the features of an igneous rock and in the field can easily be mistaken for one. The gneiss is apparently not a syntetic rock, however, since there is no evidence whatever that any appreciable amount of magmatic material was introduced during its formation. Nor is the gneiss a product of ultrametamorphism because the evidence cited previously in this paper indicates that no part of it has ever been remelted. Instead, the facts strongly suggest that the gneiss is a true metamorphic rock which has developed an igneous appearance simply by granulation, flowage, and recrystallization.

The writer is not aware that any large body of igneous-looking rock has ever been explained in such a way as this, at least not in recent literature. To be sure many igneous-looking rocks are interpreted as products of metamorphism, but in these cases the term "metamorphism" usually involves either the addition of igneous material to some older rock (syntexis), or the fusion and recrystallization of an older formation (ultrametamorphism). A fundamental, petrogenetic problem is thus suggested by the Laurel gneiss since its origin appears to be unique. It would be interesting to know whether there are other igneous-looking rocks that might have originated in the same manner.

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EXPLANATION OF PLATE 6

PHOTOMICROGRAPHS OF TYPICAL WISSAHICKON SCHIST AND LAUREL GNEISS

Figure 1.—Typical Wissahickon schist from locality 1 (Pl. 1). Foliation caused by layers of well-oriented mica alternating with layers of equidimensional quartz and plagioclase. Note straining in quartz. Diameter of field 5 millimeters, crossed nicols.

Figure 2.—Typical Laurel gneiss near locality 28 (Pl. 1). Note inequigranular texture and irregular strained quartz grains with sutured boundaries. In contrast with typical Wissahickon schist this rock has a poor foliation determined by small, crude aligned mica grains and somewhat elongate quartz crystals. Note large muscovite porphyroblast with inclusions of poorly oriented biotite. Diameter of field 5 millimeters, crossed nicols.

Figure 3.—Laurel gneiss, 500 feet west of locality 20 (Pl. 1), with large, highly strained quartz lump. Note that lump is intercrystallized with surrounding gneiss, proving it is a remnant and not introduced later. Note also the granulated, recrystallized appearance of the rock. Diameter of field 5 millimeters, crossed nicols.

Figure 4.—Photomicrograph of large quartz crystal strained and broken into fragments. The rock is from the transition zone between the Laurel gneiss and the Wissahickon formation at locality 28 (Pl. 1). Diameter of field 5 millimeters, crossed nicols.

Figure 5.—Large quartz grain shown in figure 4 of this plate. The original grain, surrounded by the heavy black line, is now broken into many fragments. Arrows and figures indicate attitude of c axes within individual fragments.

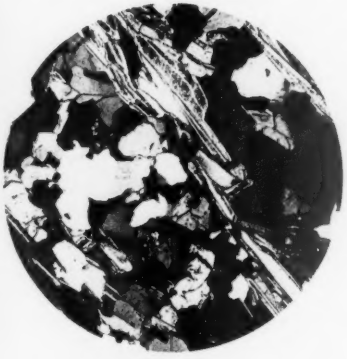

FIGURE 1

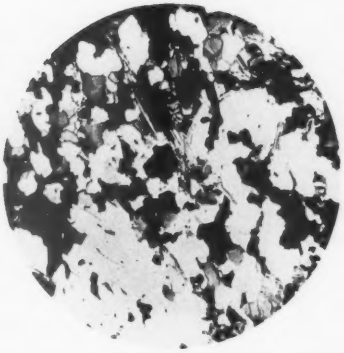

FIGURE 2

FIGURE 3

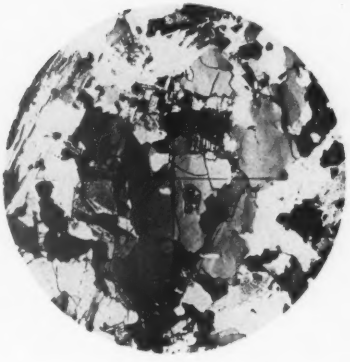

FIGURE 4

FIGURE 5

PHOTOMICROGRAPHS OF TYPICAL WISSAHICKON SCHIST AND LAUREL GNEISS

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NATIONAL RESEARCH COUNCIL AND CO-OPERATION IN GEOLOGICAL RESEARCH

BY WALTER H. BUCHER

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INTRODUCTION

All research, in the end, is the product of individual initiative and effort, so much so that we are apt to overlook the growing need for co-operation among scientific workers in the ordering and making available of results and in the planning, financing, and execution of investigations which in scope exceed the powers of the individual. The time is past when a Dana or a Zittel could collate all knowledge of one field in a handbook, nor are we inclined individually to spend "too much" of our time "compiling". And the time has come when new facts and new roads to understanding lie either in less accessible parts of the globe or in the "borderlands" of science which are no less difficult of access to the individual. Co-operation has accordingly become widespread among men in one field of science as well as across the borderlands between different sciences.

This trend toward co-operation was stimulated greatly when in 1919 [1] the National Research Council was transformed from an instrument of war to one of peace, charged with the task of promoting "research in the mathematical, physical, and biological sciences, and in the appli-

[1] The executive order is dated 1918, but the reorganization took place in 1919.

cation of these sciences to engineering, agriculture, medicine, and other useful arts. . . ." (National Research Council, 1933.) The charge was a new one as applied to the whole field of a science. How new it was is shown by the efforts that were made in each Division to define the scope of the Council's work, as illustrated in geology by E. B. Mathews' *Outline of the scope of the National Research Council Division of Geology and Geography* in the chairman's report for 1921, N. M. Fenneman's paper on the functions of the Division (1922), and the list of some 48 possible projects discussed in the first five divisional meetings (Annual Report, 1922, esp. Appendix B). Much has since been accomplished, more than is generally recognized. A brief survey of the lines of action that have proved effective would be worth while in itself now, after two decades of results. The immediate reason, however, for giving such a summary at this time lies in the growing interest of The Geological Society of America in a research "program."

The discussion that is now going on in this Society is much like that which led to the consolidation of opinion concerning a research program in the American Association of Petroleum Geologists some 14 years ago. At the meeting of the Association in San Francisco in 1928, Heald began a paper on *Research and the American Association of Petroleum Geologists* with questions such as are now being debated among members of the Geological Society, as for instance, "Should it [the Association] initiate work or should it merely support work that has been started by some other organization?" As with us now there existed then "some uncertainty as to the scope and functions of the National Research Council and the American Petroleum Institute, the two organizations that are most actively fostering research in problems of petroleum geology" (Heald, 1928, p. 939).

The first question had been answered already with firmness and vision by W. E. Wrather in the progress report of the Committee on Research, presented at the tenth meeting of the Association in 1925. He wrote:

> "In view of conditions prevailing in the oil industry, the functions of a Research Committee should be (1) to visualize and delineate particular problems of needed research; (2) to broadcast and popularize these problems to the men who are constantly in the field, and encourage them to gather and transmit information of the type needed for the prosecution of such problems; (3) to locate men who are particularly interested in these problems, and (4) to supply these men the necessary data, as well as proper facilities and financing, to enable them to make the most of the problem in hand." (Wrather, 1925, p. 2.)

This policy has prevailed and has been amply justified by results.

The uncertainty concerning the relations with the National Research Council was justified at the time. The latter's "Committee on Studies

in Petroleum Geology" was organized in the same year (1923) as the Committee on Research of the American Association of Petroleum Geologists, and all its members were also members of the Association. They were, as Heald said, "really complementary groups. They have co-operated closely ever since their inception, and there is no clear-cut line of demarcation between their functions." Several projects were undertaken jointly. When the geological map of Oklahoma was being prepared by the State Survey of Oklahoma in co-operation with the U. S. Geological Survey, the oil geologists and oil companies of the State contributed through the National Research Council (U. S. Geol. Survey, 46th Ann. Report, 1925, p. 29). When, about 1926, John D. Rockefeller placed at the disposal of the American Petroleum Institute the sum of $250,000 for research on problems in petroleum geology and the Universal Oil Products Company of Chicago donated an equal amount, the Institute turned to the National Research Council for advice and co-operation in the selection and execution of suitable projects. At joint meetings of the committees of the National Research Council and the Association, problems were discussed and projects formulated and cast into suitable form for approval by the Institute. This set a fine example of co-operation in scientific research between two national organizations. Since then the Association has become quite independent and has, in fact, set a standard in the planning and development of co-operative research projects. A close liaison in personnel and a regular exchange of information is of course being maintained, which is fruitful to both the Association and the National Research Council.

In The Geological Society of America, the idea of a "research program" is only now taking hold. It may be defined as the desire to invite the planning of major projects which the membership recognizes as yielding results of basic value to our science, to which it can point with pride and which it is therefore willing to support liberally.

The idea took form officially in 1936 when Fenneman challenged the Geological Society to think of research not only in terms of projects offered at random by individual members, but also as something that in some of its larger aspects requires deliberate collective planning for truly significant and far-reaching results. In response to his words a "Committee on Research Program" was created but was given only advisory function. In 1939 the committee still reported that "except for the work of advisory committees, a possible larger scope of initiative on the part of the Society has not been called for." In 1940 the Budget Committee suggested differentiation between normal miscellaneous projects and larger projects designed to advance geological knowledge along basic lines and tentatively earmarked the round sum of $10,000 as avail-

able for the latter. The response, both oral and in letters addressed to the chairman of the Committee on Research Program, showed that to many members the thought was still new, that they might look collectively at the needs of the science and plan deliberately to fill them by inviting and supporting investigations of large import. This habit of "looking collectively at the needs of the science" is essential for the healthy growth of a research program. That, however, has been and is the task of the Division of Geology and Geography. Its experience should be of value. It is best conveyed through a survey of the more important results of the division's labors. This is given in the following pages at the request of the Council of the Geological Society, made upon recommendation from the Society's Committee on Research Program, together with suggestions of promising lines of research that may help crystallize the concept of a research program within The Geological Society of America.

From a practical point of view, the National Research Council may be characterized concisely as (1) a forum where men of diverse affiliations are brought together to discuss the needs of the science and to arouse interest and co-operation in the development of new fields of knowledge or neglected old ones; (2) an official body that may effectively ask the co-operation of many individuals in research, and especially also in the creation of tools for research which other organizations are not interested in or not capable of creating; and (3) an organization that stands above suspicion of profit-making, in whose hands government agencies, corporations, and private individuals may place funds for use in research.

By bringing together in one organization men from all branches of natural science, unique facilities are created for promoting interdivisional co-operation which are difficult to achieve in other ways. The results of this type of work will be taken up first, not only because it includes spectacular achievements, but also because it serves to widen the angle of vision in a manner useful in retrospect and outlook along purely geological lines.

BORDERLINE FIELDS BETWEEN GEOLOGY AND OTHER SCIENCES

GEOLOGY AND PHYSICS

The largest single contribution to both physics and geology was made by the Division of Physical Sciences with the co-operation of the Division of Geology and Geography and the American Geophysical Union. In 1926, Dr. J. S. Ames as chairman of the Division of Physical Sciences set out to fill the need of systematic treatises on geophysics in the English language. The resulting series of volumes is a tribute to him and his

successors and above all to the members of the large committee to which the task was entrusted. All nine volumes of the series, *Physics of the Earth*, have appeared. The first six were published as Bulletins of the National Research Council, the others by McGraw-Hill Book Co. as follows:

I. Volcanology (77, 1931); II. The Figure of the Earth (78, 1931); III. Meteorology (79, 1931); IV. The Age of the Earth (80, 1931); V. Oceanography (85, 1932); VI. Seismology (90, 1933); VII. Internal Constitution of the Earth (1939); VIII. Terrestrial Magnetism and Electricity (1939); IX. Hydrology (1942).

In the spring of 1936 Dr. E. S. Bastin, as chairman of the Division of Geology and Geography, sent letters of inquiry to about 300 geologists and geographers asking them which lines of research were regarded as especially promising in the fields of geology and geography on the one hand and in the borderlands between geology and physics and chemistry on the other. A digest of the replies to the first question was published in 1936 under the title: *Suggestions concerning desirable lines of research in the fields of geology and geography.*[2] With it a comparable list of subjects should be consulted which was secured from Fellows of the Geological Society through a circular letter sent out by the Society's Committee on Research and published in the Proceedings for 1937 (Vaughan, 1938).

For the sifting and appraising of the replies concerning borderland problems a special, interdivisional committee was created under the chairmanship of T. S. Lovering. Its report (1938) is rich in trenchant formulations of problems and important references. It led directly to the organization of two interdivisional committees with specific functions. One was to fill the wide-felt need for a concise and up-to-date collection of the physical constants of geologic materials. Under the chairmanship of Francis Birch, this important work has been completed and published by the Geological Society (Special Paper 36, 1942). The other Committee, on Density Currents, with Herbert N. Eaton as chairman, grew out of the interest in bottom currents of "standing" water bodies which had been aroused by Daly's brilliant exposition of his hypothesis concerning the origin of submarine canyons. A large amount of quantitative data was collected in Lake Mead concerning all essential factors in the hydrology of the lake. A bulky report, containing the data in tables and graphs, has been completed and is available in typewritten

[2] For the latest list of *Publications of the National Research Council, Washington, D. C., of interest to geologists and geographers*, write the Secretary of the Division of Geology and Geography, National Research Council, 2101 Constitution Avenue, Washington, D. C.

form at a number of places in the United States, including the Division's office. A digest of the data is being prepared for publication.

In 1923, six years after Barrell had refuted Becker's ill-considered denunciation of the uranium-lead method of measuring geological time and had awakened American geologists to the possibilities of this method (Barrell, 1917), a Committee on Measurement of Geologic Time by Atomic Disintegration was organized at Lawson's suggestion with A. C. Lane as chairman. Beginning with a few pages in 1924, the committee's report quickly grew to mimeographed volumes giving annual summaries of work accomplished (annotated bibliographies) and pages of critical comments and stimulating outlook toward needed investigations. Even though it was not organized as an interdivisional committee, it comprised from its beginning the leading workers in the field of radioactivity in America and maintained regular correspondence with those in foreign countries. It has made possible much costly work through funds secured from a number of sources, including rather large sums from anonymous donors. In the later years the Geological Society has supported experimental work to the extent of more than $25,000. We have a right to be proud of what has been accomplished under our auspices. Few realize as yet in how many directions the newest development in this field, Goodman's and Hurley's work in the helium-content of heavy minerals, promises to shed new light. The utilization of the radioactive record for geological purposes actually has barely begun. The Advisory Committee on Research Program, accordingly, recommended last October to the Council of the Geological Society "that investigations of the radioactive measure of geological time be continued as a part of the Society's program which is given major emphasis."

The support which the Society has given to this work justifies reference to it as part of the Society's program. It illustrates one of several ways in which the units of a research program come into being.

It is to be expected that, in general, units of this program which require for their formulation the co-operation of organization physicists and geologists should take form and find impetus in the American Geophysical Union. This organization was established in 1919 as the American Committee of the International Union of Geodesy and Geophysics under the executive control of the Committee on Geophysics of the National Research Council.

"The objects of the Union are to promote the study of problems concerned with the figure and physics of the Earth, to initiate and coordinate researches which depend upon international and national cooperation, and to provide for their scientific discussion and publication."

The growth of the "Transactions" through the years offers a remarkable example of the good that comes from bringing together men from two fields of science who not only speak the same language but whose intellectual curiosity extends far enough to embrace the other's range of vision.

The records of the Division of Geology and Geography contain more than one example to show that without team work across the borderline the geologist can achieve little along geophysical lines of inquiry. In 1923, for example, a "Committee on Detection of Subsurface Structures by Sounding" was discontinued because the circumstances under which investigations in this subject were conducted at the Bureau of Standards made it impossible for the committee to procure information at that time. The geologists apparently were not then thinking of other sources of information and co-operation.

A "Committee on Sonic Sounding and Oceanographic Thermographs" was organized in 1922 at the suggestion and under the chairmanship of T. Wayland Vaughan. It was renamed in 1925 "Committee on Submarine Configuration and Oceanic Circulation."

"The principal duty of this committee was to obtain information for distribution to geologists and geographers, and also to put the members of the Division in a position to lend support to investigations of the two kinds mentioned." (Annual Report, 1923, Appendix "H".)

At a time when few American geologists knew anything about investigations in physical oceanography, and years before the International Council for the Exploration of the Sea began to publish such information for oceanographers, this committee did most valuable service through its annual reports, the last two of which comprised over 100 pages each. But it is significant that among the duties of the committee no mention is made of deliberate planning of work designed to secure knowledge of basic value to geology and enlisting the co-operation of competent individuals and organizations in its execution. Such co-operation in planning and doing requires that the right men be brought together. Chiefly to make this possible, the functions of this committee were transferred in 1932 to the Geophysical Union. In this new environment such investigations as Ewing's seismic studies on the structure of the continental shelf and Piggot's and Stetson's studies on Atlantic sediments were vitalized if not conceived. (*See, e.g.*, Am. Geophys. Union, 1932.) The same may be said, *e.g.*, of Woollard's magnetic and gravitational surveys; Birch's and Slichter's studies on the velocity and propagation of seismic waves; and Grigg's work in mechanics of rock deformation under high pressures and temperatures.

The energy that is needed for such studies is drawn from two soils: that of geophysics which provides the methods of geodesy, oceanography, seismology, and physics of materials; and that of geology which provides the comprehensive body of knowledge from which the demand for new information springs. The work is done by men who are members of our Society. Their investigations are ours and fit beautifully into the requirements of a "research program" as stated above:

"major projects which the membership recognizes as yielding results of basic value to our Science, to which it can point with pride, and which it is therefore willing to support liberally."

GEOLOGY AND CHEMISTRY

Geochemical problems have been less in the limelight in the last two decades than those of geophysics. Many have been under investigation for decades in the Geophysical Laboratory. Some that are no less pressing are waiting to be studied. Fleischer and Morey's work on the *Equilibrium between vapor and liquid phases in the system* CO_2-H_2O-K_2O-SiO_2 is a good example of an important investigation that was initiated by a small group within The Geological Society of America. The problem and its method of attack were formulated in 1935 at conferences between A. L. Day, G. W. Morey, C. H. Behre, Jr., and T. S. Lovering. The work was carried on in the Geophysical Laboratory. It was financed for two years by grants from the Penrose Bequest and for a third year by the Carnegie Institution of Washington. The results were published in 1940.

The report of Lovering's Interdivisional Committee previously referred to contains valuable suggestions. Since its publication, steps have been taken to secure action along some lines.

At a two-day meeting in October, 1940, called for the purpose, the Committee on Problems of Ore Deposits, with Lovering as chairman, analyzed in detail the whole range of problems connected with the physical chemistry of ore deposits. The "Summary of problems suggested" as a result of the joint counseling of five chemists, one physicist, and ten geologists is an important document (Annual Report 1940-1942, Appendix J). The committee went further, however. Having formulated the problems, it proceeded to decide on what to begin in a systematic series of investigations designed to supply vital basic data. The water-sulphide systems proved both feasible and basic. A first unit in this planned series was chosen: the study of the solubility of metallic sulphides in water at elevated temperatures and pressures. As before, the Geophysical Laboratory has consented to provide space, light, and power for the research worker, and, above all, its unique facilities and

the invaluable advice of its staff. The Geological Society has agreed to finance this unit of the investigation.

GEOLOGY AND BIOLOGY

In contrast with this working hand-in-hand of geologists with physicists and chemists, the lack of systematic joint counseling of paleontologists with biologists is conspicuous. There have been no interdivisional committees, either of Geology or of Geography and Biology, in the National Research Council. Within the division of Geology and Geography, two committees on Paleoecology (1935-1937 and 1941) made a good start, however. The first of them included three zoologists: Paul Bartsch, Austin H. Clark, and M. W. de Laubenfels. In his first preliminary report the chairman, W. H. Twenhofel, said explicitly that the committee "was not organized to undertake research in the field of its study." Its object was "assemblage of data bearing on the ecologic or environmental relation of fossil organisms. . . ." The work of the committee was understood to be "in the nature of propaganda to bring the subject of paleoecology to the attention of students so that it may be talked about and become a part of the general thinking." This sort of service is as far-reaching as its effects are hard to trace. It may be said to be the first function of every organization that undertakes to stimulate research.

The work of this committee recalls that of the Committee on Submarine Configuration and Oceanic Circulation. But unlike the latter, it terminated its own existence after issuing but two reports (1936, 64 pages; 1937, 63 pages). Of individual groups, it had treated of vertebrates, arthropods in general and trilobites in particular, Paleozoic cephalopods, brachiopods, Porifera, and Foraminifera. Of environments, the Lower Part of Florida (with special emphasis on mollusks) was the only one covered. The Committee had obviously barely scratched the surface of this vast subject. Yet the student of sediments, as well as the paleontologist, urgently needs an understanding along the lines of ecology. Why was there no outcry of protest from workers in these fields when this valuable series stopped abruptly? Did they know about it at all? The division of Geology and Geography is taking the latter question seriously, as applied to all its products. There is good evidence that our division has been hiding its light under the proverbial bushel. Steps have been taken to make more widely known the first report of the new "Committee on Ecology of Marine Organisms" (Annual Report 1940-1941, Appendix A-1, 52 pages).

The new name of the second Committee reflects a change in the outlook of its personnel, which comprises chiefly men interested in the physiologic-oceanographic aspects of ecology, in the probing into the

factors that control the character and distribution of life. The morphologic-zoogeographic-paleontologic element which prevailed in the first committee might well form a complementary group devoted to the presenting of the results produced by physico-chemical factors of the environment within the framework of space and time.

It is to be hoped that the present committee will eventually take the step which so greatly vitalized the Committee on Submarine Configuration and Oceanic Circulation. There is, of course, no "geobiological" counterpart to the Geophysical Union, but the end can be achieved by bringing together two types of men: one, geologists and paleontologists who recognize definite groups of facts which will be understood only when certain gaps in the knowledge concerning modern organisms have been filled; and the other, the biologists and oceanographers, who have the techniques and the interests required to fill those gaps. Out of this meeting of minds comes deliberate planning and directed research and from it that awakening of mutual interest in and respect for the other's field which is necessary for a deepening of understanding of problems and realities on both sides.[3]

But ecology is only one aspect of biology. Is there no need for getting together in other fields? "Evolution" is still full of riddles. The disciples of the young and vigorous experimental science of genetics have little time for, and less interest in, the body of solid fact which the older, undaunted historical disciplines of morphology, taxonomy, and paleontology are accumulating. Insofar as the disciples of the latter exhibit interest in genetics, it is mingled with distrust in the long-range significance of the former's findings. Here is a field that requires joint counseling. The new and independently created informal "Society for the Study of Speciation,"[4] the "dynamics of the origin of species," will go far in this direction. "The general object of the Society is to institute an informal information service which will tend to correlate the various approaches." The Paleontological Society was quick to invite its members to join. It is represented on the Executive Committee by Dr. G. G. Simpson, a leader among the vertebrate paleontologists who individually and as a group have kept most directly in touch with modern developments in biology. May we look to them to take the initiative in bringing together their own men with invertebrate and plant paleontologists for an enlivening discussion of "the dynamics of the origin of species," perhaps under the auspices of the National Research Council?

[3] Similar developments are to be hoped for in the other fields of ecology, those of fresh-water and land environments.

[4] For information address the Secretary, Alfred E. Emerson, Department of Zoology, The University of Chicago, Chicago, Illinois.

And how about zoogeography, past and present? The hypothesis of the drifting of continents was conceived and developed by the meteorologist Wegener and supported dominantly by zoogeographical observations. His hypothesis must have aroused interest in this fascinating field of knowledge among geologists and paleontologists. Paleobotanists have long cultivated energetically the field of phytogeography. Zoology, botany, and geology would benefit from activities that would open up the newest advances as well as the basic principles of zoogeography and phytogeography to a wider audience.

Even the lowly field of "ichnology," the study of tracks, trails, and burrows left behind by organisms, holds important keys to geological problems and deserves to be brought to more general attention. It has recently demonstrated the presence of King crabs (limulids) in the Triassic of New Jersey (Caster, 1939) and has disproved what for three-quarters of a century had been interpreted as evidence of frequent exposure to the air of the lime oozes of the Upper Jurassic embayment of the classic Solnhofen region (Caster, 1940). It should increase our understanding of pre-Cambrian life materially, if an interested group of workers made an intensive search for tracks and trails of what college texts still call "worms." The "very numerous" tracks of mammals, including small elephant and modern-looking canid, horse, and artiodactyl tracks, probably of Pliocene age, which Curry discovered on the east side of Death Valley (Noble, 1941, p. 956), are of great interest beyond paleontology. They may represent a relatively recent counterpart to the famous dinosaur tracks. If studied by men familiar with techniques of sedimentation, climatology, geomorphology, and mammal biology, they may help solve more precisely the question of the climatologic and topographic environment during the Triassic in the region of the present Connecticut Valley.

GEOLOGY PROPER

GENERAL STATEMENT

Turning from the borderline fields to geology proper, we recognize the same trend of development, from first efforts to create bibliographic tools and spread knowledge of, and interest in new or neglected lines of study, to a critical synthesis of knowledge that leads to co-ordinated investigations focused by the newly gained perspective. It is convenient to report the examples of these activities under corresponding headings.

BIBLIOGRAPHIC TOOLS AND REPORTS OF PROGRESS

Among the earliest publications of the Division of Geology and Geography two bibliographic lists stand out: E. B. Mathews' (1923) *Catalogue of published bibliographies in geology, 1896-1920* and H. P. Little's (1922)

List of manuscript bibliographies in geology and geography. Bibliographies or annotated lists of papers on specific subjects were contributed in separate issues by a number of committees as an effective way to open up a new or newly active field to geologists at large, as, for instance, isostasy (Adolph Knopf, 1924), the mechanics of igneous intrusion (F. F. Grout, 1935), and structural petrology (T. S. Lovering, 1938).

A contribution of major proportions, envisaged in one form or other since the earliest days of the National Research Council, became reality when under the leadership of Waldemar Lindgren a committee succeeded in raising a total of $21,358.58 from mining and oil companies and individual donors and thereby made possible the publication of the *Annotated bibliography of Economic Geology.* The first volume, for the year 1928, appeared in 1929, under the sponsorship of the National Research Council. In 1936 the sponsorship was transferred to the Society of Economic Geologists. For several years now the Geological Society has supported it liberally.

While speaking of bibliographies we may well point with pride to the *Bibliography and index of geology exclusive of North America.* With this publication added to the U. S. Geological Survey's *Bibliography of North American geology,* the first and only international bibliography of geology has been created in the English tongue. It has been made possible through the close and harmonious co-operation between the U. S. Geological Survey and the Geological Society.

Two committees have set unique records: the Committees on Paleobotany and on Sedimentation. The Committee on Paleobotany was among the first to be created in the Division of Geology and Geography (1919). Beginning in 1921,[5] it has reported annually, with the exception of an early gap of several years, the important papers published in North America during the year, together with comments on work in progress. Bound together, these reports represent a historical record of real value. Relatively few geologists and even paleontologists know about it. It deserves thoughtful consideration. Would not every unit in the large field of geology benefit by such an annual summary? Suppose we had a similar annual record of work accomplished, let us say on the structure of the Appalachians or the Great Basin, or on the mechanics of solids in its application to structural geology, or in petrology. Such a record would be no more difficult to create and would be no less valuable. In fact, if sufficiently condensed and cast into the right form, such annual reports might enable many workers in different fields to keep in touch with the essentials of progress along diverse lines of

[5] The first report appeared in 1923. But, in addition to the report on the current year it contained summaries for 1921 and 1922 by David White and E. W. Berry respectively.

geology too far removed from their own. Such a work, clearly conceived and effectively carried out, would be an excellent antidote to the constriction of vision which accompanies all specialization.[6]

Perhaps the most remarkable record has been set by the Committee on Sedimentation which, since its inception, has pursued vigorously a number of lines of usefulness. It has developed the policy of publishing progress reports (with bibliographies) on a given subject at unspecified intervals of a number of years. Each annual report contains several such progress reports. In volume, its annual reports[7] rose quickly from 5 pages in 1920 to 68 pages in 1923 and have averaged close to 100 pages since then. In addition to reports of progress, they contain papers concerned with the standardization of nomenclature and laboratory methods, and lately especially also summaries of new knowledge in allied fields such as hydrology and colloid chemistry. To secure greater completeness and precision in the description of sedimentary rocks, the committee prepared and had published a *Schedule for field description of sedimentary rocks* (M. I. Goldman and D. F. Hewett, 1923) and a *Color chart for the description of sedimentary rocks* (M. I. Goldman and H. E. Merwin, 1928). Two books have carried the direct influence of the work of this vigorous committee into every geological library and classroom in the English-speaking world: a symposium welded into a coherent work by W. H. Twenhofel in *Treatise on sedimentation* (1st edition, 1926; 2nd revised edition, 1932) and the symposium edited by Parker D. Trask in *Recent marine sediments* (Am. Assn. Petrol. Geol., 1939). These exceptionally influential works, like the committee's Annual Reports, contain in a large measure the sort of synthesis that creates perspective and reveals gaps in our knowledge. Work of that character is treated more specifically under the next heading.

CRITICAL SYNTHESIS OF DATA, ANALYSIS OF PROBLEMS, AND ORIGINAL INVESTIGATIONS

When the newly created Committee on Tectonics held its first meeting in December 1922, it voted "to commence work on a tectonic map of North America which should bring out the trends of folding, the principal lines of faulting, the axes of doming and related structural features." The immediate impulse for this ambitious undertaking had come from the epochal structure map of Eurasia which Argand had exhibited that sum-

[6] *Compare* Geologische Jahresberichte, begun 1938. (Gebr. Borntraeger, Berlin.)

[7] It is unfortunate that these reports were not published in uniform format and print. Reports 1 to 3 (1920-1922) were issued mimeographed and unbound; 4 to 8 (1922-1928) paper-bound mimeographed; 9 to 11 (1928-1930) printed in the Reprint and Circular Series as Nos. 85, 92, 98; Nos. 12 and 13 (1930-1934) printed as Bulletins 89 and 98 respectively; Nos. 14-18 (1936-1940) were again issued in paper-bound mimeographed form.

mer at the meeting of the International Geological Congress at Brussels.[8] It could not be said to have grown from a specific desire on the part of American geologists for such a map. In fact, the word "tectonic map" did not convey much, nor the same idea, to most men. So the plan went through a period of incubation. "The time was not ripe," *i. e.*, we geologists were not ripe for this undertaking. When plans were made for the International Geological Congress to meet at Washington, D. C., G. R. Mansfield encouraged Philip B. King of the U. S. Geological Survey to prepare a general description of major structural units in the United States, accompanied by a small-scale tectonic map (published as Guidebook 28, 1932). A little later, Ralph D. Reed, in co-operation with J. S. Hollister, undertook the preparation of a tectonic map of Southern California which set a new standard and attracted wide attention (Reed and Hollister, 1936). These two publications represent high points on a rising wave of general interest. Under its impact, in 1935, under the leadership of Chester R. Longwell, the construction of a tectonic map was begun in earnest by the Committee on Tectonics. It was restricted to the United States and planned on a larger scale—that of the new geologic map of the United States. The compilation was undertaken by eleven subcommittees (Nat. Res. Council, Div. Geol. Geogr., Ann. Report for 1934-1935, Appendix D). Three years later, with the financial help of the American Association of Petroleum Geologists and The Geological Society of America, the maps furnished by the subcommittees were assembled in a first tentative draft. Two hundred black and white lithoprinted copies of this map were distributed among organizations and individuals for corrections and additions. The response was gratifying. The effective co-operation of the American Association of Petroleum Geologists and allied local societies deserves special praise. The final revision and drafting of the map is now in the able hands of P. B. King and will be completed as quickly as the present emergency permits. Arrangements for publication have been made with the American Association of Petroleum Geologists,

[8] While speaking of the results of co-operation in geology we do well to recognize that we also owe to the catalytic action of the International Geologic Congress our two most important summary maps: (1) The creation of a geological map of North America (1:5,000,000) was proposed by J. C. Russell at the Ottawa meeting of The Geological Society of America (Dec. 1905) in connection with the impending meeting of the International Congress at Mexico City in 1906 (G.S.A., Bull., vol. 17, 1906, p. 709; Cong. Geol. Int. 1906, C.R., 1907, p. 211). It was produced, apparently in record time, under Bailey Willis' direction through the co-operation of the Surveys of Canada, the United States, and Mexico. A preliminary edition was published in 1906 at the expense of the Congress (U.S.G.S., 27th Ann. Report, 1906, p. 21); (2) In that same year (1906) plans concerning a geologic map of the United States on the scale 1:2,500,000 are mentioned in the Annual Report of the U. S. Geological Survey. For many parts of the country, however, not enough detail was available at that time for a map on so large a scale. Twenty years later, when the time was drawing near again for an invitation to the United States—in the winter of 1926-1927—following the meeting of the International Congress in Spain, the compilation of such a map was begun in earnest (U. S. Geol. Survey, 48th Ann. Report, 1927, p. 71). It was ready for distribution when the Congress met in Washington in 1933.

one of three organizations that had expressed interest in the publishing of the map.

As in other cases, development of this project is sketched here in some detail because it illustrates well a number of matters that concern everyone interested in the development of a research program.

The story of the glacial map of North America is equally instructive. A committee was organized by the division in October 1939, under the leadership of R. F. Flint. In contrast to the tectonic map, the time was overripe for this kind of a map. There was no period of incubation. Work progressed rapidly. Gaps were filled by field work, some financed by the Geological Society. A symposium was held at the Boston meeting of the Society in 1941. The National Research Council division allotted a total of $1000 for the traveling expenses of committee members in order to facilitate the working out of the details for the first draft. The American Geographical Society contributed $1000 to secure the services of an experienced draftswoman for D. A. Nichols of the Geological Survey of Canada, who is giving practically his full time to the work of this committee. From the start, the Canadian Survey has set a fine example of official co-operation. The map is being drafted on a new base which is in process of being computed and plotted. The map will contain a great deal of new and unpublished information derived from the recent exploratory work of the Canadian Survey and numerous other sources.

The glacial map of North America will be published on the scale of 60 miles to the inch. It will be provided with submarine contours from data furnished for Canada by Nichols, for the United States by the Coast and Geodetic Survey through the interest of Lieutenant Paul A. Smith. The Geological Society is looking forward to publishing it.

These two maps, tectonic and glacial, when completed will be a challenge to American geologists. They will be criticized for some of the things shown on them and more for what is not shown. But, above all, they will be used and they will lead to new thought, new work, new knowledge, all on a larger perspective than has been possible without such attempts at synthesis.

The new correlation charts of the Committee on Stratigraphy perform a similar service. This committee does not deal with the shape and structure of sharply circumscribed bodies, but with the relative position on the time scale of larger and less easily definable categories, the formations. Both, the definition of these units and their relative incidence in time, are in many cases matters of judgment rather than of direct observation. It is, therefore, difficult to achieve agreement among different workers. Although begun in 1932, this work has not been completed yet.

Two of the twelve charts and a general introductory statement have appeared in the March and April issues of the Society's Bulletin for 1942; five others are virtually ready for the printer, and work on the remaining five is in progress.

A fine example of a sharply focused, broadly co-operative enterprise of a different kind is that which finally resulted in the publication of the Society's Special Paper No. 24, Dec., 1939: *Contributions to a knowledge of the lead and zinc deposits of the Mississippi Valley region*. This work was produced under the leadership of E. S. Bastin and edited by him. It began in 1932 with the formation of a subcommittee of the Committee on Ore Deposition. In the first two years, the committee met three times—at St. Louis, Mo., Chicago, Ill., and Baxter Springs, Kansas. A total of 45 people attended the last meeting, at which papers were read and excursions held. Graduate students at four universities and most of the members of the committee undertook special mineralogic, petrographic, and chemical studies. Stratigraphic and structural data were collected. At the end of five years, when the material was practically ready for the press, the committee asked to be discontinued, having secured a noteworthy addition of knowledge to a basic problem of ore genesis. (*See* Annual Reports of the Division from 1933 to 1937.)

In 1936, when W. H. Newhouse became chairman, the Committee on Processes of Ore Deposition took steps to prepare a symposium on the influence of structure in localizing mineralization. Over 70 manuscripts have been edited and combined in a large volume which is being published by the Princeton University Press.

It is, of course, impossible to speak here of the work of all the committees of the Division, through the last 20 years. Two further examples illustrate still different ways of securing results. Grout's Committee on Batholith Problems, which functioned between 1928 and 1935, issued two thought-provoking reports: *Problems of batholiths* (rev. ed., 1933) and *Comments on magmatic stoping* (1935), the former a broad analysis, the latter chiefly a collection of quotations.

The work of the Committee on Shoreline Investigations has produced two unique results. Studies on shore phenomena, in co-operation with the U. S. Coast and Geodetic Survey, were among the first suggestions to be made in 1919 to the Division of Geology and Geography of the newly created National Research Council. In 1923 a Committee on Shoreline Investigations was appointed with Douglas Johnson as chairman. This committee undertook two projects: first, a specific one, *viz.*—to test the theory of local variations in mean sea level. The New York waters were chosen for the test. The U. S. Coast and Geodetic Survey, the Department of Docks of New York City, and the Department of

Plant and Structure of New York City co-operated with the National Research Council. The results of this investigation were published by Douglas Johnson in Bulletin 70 of the National Research Council (1929).

The other project was a very broad one: to stimulate interest in the systematic collection of data concerning changes in the beaches of the Atlantic seaboard which might later be studied "collectively, to deduce the underlying principles, and separately, to ascertain the variations from these principles to which individual beaches are subject." (Ann. Report, 1927, App. F, p. 3.) In order to create interest in this subject, with its wide social and economic implications, a circular letter was sent in 1926 to some two hundred addresses along the Atlantic and Gulf coasts, stating the Committee's

> "conception of the problem, making certain specific inquiries regarding conditions along the different sections of the coast, and requesting an expression of opinion as to the feasibility of securing public support for the cooperative study which the committee had in mind."

Ensuing correspondence, contact with officials of New Jersey, and resulting conferences led eventually to the formation, in the same year, of the American Shore and Beach Preservation Association. This active organization, in turn, has been a factor leading to the formation of the Beach Erosion Board in the Corps of Engineers of the United States Army. Although not directly connected with the Research Council, the American Shore and Beach Preservation Association usually holds its annual winter meeting in the National Academy-Research Council Building. In 1933 it started publication of a new journal "Shore and Beach." The work of the Division's committee which was discontinued in 1935 is thus being carried on on a broad basis.

OUTLOOK

Along purely geological lines no less than in the borderlands large problems invite joint planning and action. Let us round out this discussion by examining a few.

Regional studies, especially of major structural units, suggest themselves at once. In fact, the systematic study of the Rocky Mountain Region was one of the very first projects suggested to the newly formed National Research Council (Blackwelder, May 1, 1919). Such studies have been organized successfully by members of The Geological Society of America with the support of university departments, as, *e. g.*, Kellum and Imlay's studies in the folded belt of northern Mexico [9] and the work

[9] This has resulted so far in the publication of one book (by the University of Michigan), about 10 papers that have appeared in the Bulletin of The Geological Society of America, and one printed by the American Association of Petroleum Geologists.

of the Yellowstone-Bighorn Research Association.[10] Neither of these was undertaken deliberately as part of a program by the Geological Society as such, nor were they financed exclusively by it, though both received liberal assistance.[11]

Where important gaps exist in our knowledge of critical regions, planning will yield important results within our own borders and, more so even, in the vast regions of Latin America. The obstacle that blocks attempts at larger planning in such directions is a curious, indiscriminate fear of areal mapping, as if regional structures could ever be grasped without it. No one would propose random mapping for mapping's sake, nor is it our function to extend further any survey into territories that may not be expected to shed new light on some larger question. But we cannot afford to frown on an investigation because it involves areal mapping as an essential approach to a vital problem of structure.

There are, of course, large regional problems which require other methods for their solution. One concerns the synchronism and the extent of mid-Pleistocene continental deformation. It has been suggested by the writer that most of the United States and perhaps this and other continents as well have experienced a major epeirogenic uplift during mid-Pleistocene time which was responsible for the rejuvenation ("deep valley" stage) of stream valleys in the mid-continent regions as well as in those of the Cordilleran Mountains and which may have been contemporaneous with the mid-Pleistocene folding of the Californian Coast Ranges (Bucher, 1940, p. 422, footnote). This is a large problem that can be solved only by geologic methods and actually goes as deep into the problem of crustal deformation as any investigations by geophysical methods can hope to do. It is the sort of problem that should be attacked in an organized, co-ordinated effort. Here is material for a future major project of the National Research Council or the Geological Society or both.

Closely allied with it is another question of continent-wide interest. From a cross section of the terraces in the Susquehanna River valley, M. R. Campbell has attempted to compute the relative volumes of rock that were removed from the valley in the intervals between successive uplifts (Campbell, 1933, p. 572). Would it not be of interest to know what these ratios are for typical streams all over the country? Here is another project calling for co-operative, country-wide study from which valuable insight into time relations of epeirogenic processes will result.

The last project serves to emphasize the general need of quantitative data concerning geological processes. We need a companion volume to

[10] This has led so far to the publication of about 85 scientific papers in different periodicals.

[11] Kellum's work also received a grant-in-aid from the National Research Council.

Birch's *Handbook of physical constants,* one giving reliable quantitative data concerning rates involved in all sorts of geological processes and the factors that control them.[12] This leads straight to the borderland between geology and applied physics such as soil erosion, hydrology of running water and shore lines, and meteorology.

The manner in which the stratigraphic timing of strata ("correlation") is recorded is highly unsatisfactory. The writer has discussed this matter in a paper given at the Boston meeting entitled *Method proposed to introduce the concept of "limits of error" into the stratigraphic timing of tectonic movements.*[13] He believes that improvements along the lines suggested by him should be attempted. This can best be done by a concerted joint effort.

A unique opportunity for studies bearing on the major problems of crustal deformation is offered by the photographic record of the Ross Shelf Ice which has been secured by Byrd's Antarctic Expeditions. Locally, as, *e. g.,* in the Bay of Whales, the ice is thrown into systems of folds or is traversed by large rifts or cut by systems of fractures (Gould, 1935). Much of the ice rests on a rock shelf, some of it floats. Most of it is in movement. A committee of the Division of Geology and Geography of the National Research Council, just organized, has undertaken to prepare a tectonic map of the Ross Shelf Ice and to attempt an analysis of the mechanics that determine the nature and distribution of the structures. It is hoped that this huge laboratory experiment of nature will yield new clues for an understanding of the deformation of the earth's crust.

Fringing the borders of the continents, the marine terraces offer a problem parallel to that of the river terraces mentioned above and directly related to it. About 10 years ago, Douglas Johnson submitted a project to the Geological Society which provided for a series of co-ordinated, simultaneous studies of typical sequences of terraces on different shores along the Atlantic Ocean. Like another broadly conceived plan that was submitted by A. C. Veatch about the same time, this project was shelved at the time for various reasons. But as our vision and experience grow and we learn to think in larger terms, we shall return to this kind of investigation. Others have pointed to the example of the American Geographical Society which with its superb Hispanic American mapping program has set a unique example of courage and vision in long-range planning.

It is not worth while to prolong this looking ahead in the field of

[12] *See, e.g.,* Fenneman's comments on *The actual rates at which rocks waste away* (Fenneman, 1938, p. 146).

[13] Publication deferred.

geology. For a veritable treasure of "Suggestions concerning desirable lines of research in the field of geology and geography" the reader is referred once more to the report of that title which was issued by the Division in 1906 under the editorship of E. S. Bastin for physical geology, C. O. Dunbar for Paleontology and Stratigraphy, and R. S. Platt for Geography.

One item barely touched upon in that report deserves being set forth in some detail. It will close this discussion fittingly.[14]

The modern methods of echo sounding have at last brought within our reach definite information concerning the geomorphology of the vast spaces covered by the ocean floors. Many echo-sounding profiles are accumulating in the files of the Coast and Geodetic Survey and the Hydrographic Office of the Navy. A year ago the Committee on Research of the Geological Society directed the Council's attention to the value of these records. It is necessary that steps be taken to utilize them before the impression gains ground that further additions of new material and even the storage of the old represent waste.

These profiles give a picture of the configuration of the sea floor, not only in two dimensions, but in three, provided a large number of parallel and intersecting profiles is used. Unfortunately they tell little at first sight, and they must be used with full appreciation of all possible sources of error. The mere publication of a few selected profiles or hasty generalizations by one not thoroughly familiar with their use would almost certainly do harm.

We need men who deliberately train themselves for a field that is almost wholly new. To be successful in this field a man must have a thorough grasp of the problems of modern tectonics and geomorphology. He must first gain experience in three-dimensional analysis of surface forms based on profiles of greatly exaggerated scale. The best results, for a long time to come, will be attained in submarine geomorphology from the analysis of the profiles themselves. But contouring—on maps of relatively small scale—will have its place. Both methods presuppose great care in securing the most precise location possible for all the data on maps.

This means that in addition to a thorough geologic-geomorphologic training, the novice in this field should become familiar with the methods of echo sounding and the problems of position finding. This can be achieved best by studying the methods and practice of the Coast

[14] The Subcommittee on Ocean Bottoms of the Geological Society's Committee on Research Program was directed by the Council (October, 1941) to call the attention of the membership to the promise which the relatively new field of suboceanic geomorphology holds. The subcommittee has decided to embody in this address the following statement prepared by the chairman of the subcommittee, Lieut. Paul A. Smith, and the writer.

and Geodetic Survey. That organization, in times of peace, is usually seeking at least one or two adequately prepared and capable young men for various phases of survey and cartographic work.

Such work, by the way, would give the student an opportunity to get acquainted with an occupation which he may later choose as his life work, if he so desires. If, however, he prefers to stay in geology, he should have little difficulty in finding a place in the academic world, provided our universities keep their doors open to men whose minds travel new roads.

CONCLUSION

With this vision of a way to grasp the geomorphology of the larger part of our globe that lies concealed beneath the sea, this discussion of research in geology and the National Research Council comes to an end. The story of co-operation in research embraces all our geological organizations. There is no clear line of demarcation between a "research program" of The Geological Society of America and that of the National Research Council. Whether there shall be one or not does not concern us here. Whatever is done officially, the greatest possible freedom should be permitted for the formulation of projects to suit the whims, prejudices, and jealousies of individuals or groups to which science is an unwilling heir. If this manner of growth appears less orderly than one might wish, it is also a secret and great source of strength. It is the very essence of democracy, as long as we stand ready to welcome and support worthy work from whatever source it may spring.

POSTSCRIPT

The essence of this paper was given orally at the Boston meeting of the Society, only a few weeks after Pearl Harbor. Since then our outlook has been turned rapidly to the cruel needs of the hour. The words "research" and "co-operation" have taken on a harder meaning. The offices of the National Research Council are rapidly being geared into the nation's war work, and members of the Geological Society are leaving their posts in increasing numbers to go into active service of one kind or another. Already the preoccupation with problems of pure geology, pursued without thought of power or possession, begins to seem remote and perhaps almost futile.

Let us not be misled. No matter how long it may last, the war will be over soon when reckoned by generations. The material destruction can and will be repaired. It is imperative that we keep our minds turned firmly to that future which justifies today's sacrifices, when those of us who survive and those who will follow our steps will return to the

laboratories and studies. To them the preceding pages are addressed above all, to be read when they again take up the threads of thought and resume the creative dreams of free inquiry into nature.

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FLOOD DEPOSITS OF ARROYO SECO, LOS ANGELES COUNTY, CALIFORNIA

BY W. C. KRUMBEIN

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ABSTRACT

This study represents a continuation of earlier work in San Gabriel Canyon. Deposits left in Arroyo Seco by the flood of March 2, 1938, were mapped in the field, and samples were taken to determine how sedimentary characteristics varied with distance from source. In an area of Lowe granodiorite near the headwaters of the stream size and roundness of particular pebbles, free from contamination, were studied. As in San Gabriel Canyon, the flood deposits were arranged largely in log and boulder jam deposits, which formed a series of partially overlapping features along the valley. These jam deposits have reverse size gradients,—mean size of the boulders increased markedly downstream.

In general, however, neither the size nor the sphericity of pebbles in the flood deposits, collected in corresponding portions of successive log jams, showed any change along the valley; roundness, on the other hand, increased noticeably downstream. These results agree with findings in San Gabriel Canyon and suggest that size and shape respond to the dynamical conditions at the immediate site of deposition, whereas roundness is a function of the abrasion suffered by the particles over the distance traveled.

An attempt is made, in the light of available evidence, to reconstruct the physical conditions governing the movement and deposition of the flood material, as well as to evaluate the relative roles of wear and selective transportation during floods.

INTRODUCTION

This paper continues the study of deposits associated with the flood in the Los Angeles region on March 2, 1938. In an earlier paper (Krumbein, 1940) the flood deposits of San Gabriel Canyon were described, based largely on a laboratory study of selected samples. This later study was done during the summer of 1940.

In many of the canyons the flood deposits either were not adequately exposed or they had been removed in part by road building and other operations. Conditions in Arroyo Seco, however, were almost ideal. This canyon had been closed to the public since the flood, and in June 1940 the deposits were still fairly continuous and well exposed. Six weeks were spent in a detailed study of parts of the canyon. Field evidence was supplemented by laboratory studies on samples, as well as by data obtained from engineers and others who had studied the effects of the flood.

Additional work was done in San Gabriel Canyon to complete details and test tentative conclusions. In general, the new work supports previous findings, although some modification of detail was necessary.

More time and a closer field examination of the deposits in another canyon affected by the same flood permitted the further testing of some tentative conclusions of the San Gabriel Canyon study and a better classification of flood deposits. The experience gained permits at least qualitative conclusions regarding the mechanics of debris movement and deposition by floods, as well as a statement of the problem of wear and selective transport along the canyon.

ACKNOWLEDGMENTS

This project was made possible by a grant from the Penrose Bequest of The Geological Society of America. The writer also wishes to acknowledge his indebtedness to numerous individuals who aided in the work. He is particularly indebted to Mr. Lyman Huff, a graduate student at the University of Chicago, for able assistance in the field and laboratory throughout the summer. Dr. E. Wayne Galliher (Barnsdall Oil Company), Dr. Ian Campbell (California Institute of Technology), Dr. James Gilluly (University of California at Los Angeles), Dr. Vito Vanoni (U. S. Soil Conservation Service, Pasadena), and Mr. D. M. Ilch (U. S. Forest Service, Glendora) were particularly helpful in their suggestions and criticisms of the work. Mr. M. F. Burke of the Los Angeles County Flood Control furnished much valuable information regarding the engineering problems associated with the flood. The Department of Geology of the California Institute of Technology kindly furnished office facilities, and the Soil Conservation Service extended the privileges of its Fineness Laboratory for mechanical analyses.

GEOLOGICAL SETTING OF ARROYO SECO

Arroyo Seco rises in the San Gabriel Mountains and flows generally southwest and south to its junction with the Los Angeles River within the city limits (Fig. 1). The area studied extends from the headwaters to the mouth of the canyon. The rectangle in the upper right of Figure

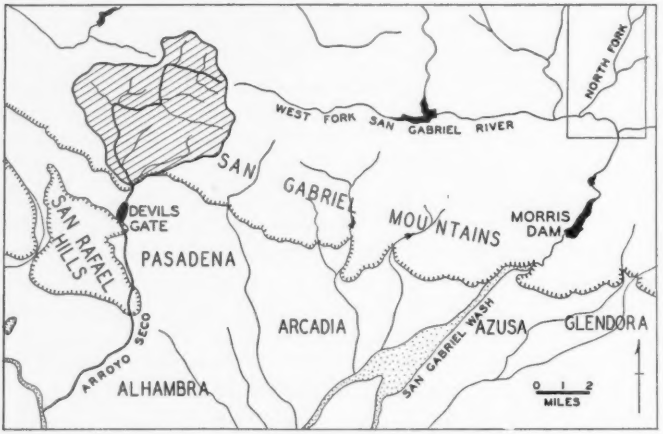

FIGURE 1.—*Index map of area studied*

The shaded area is the drainage basin of Arroyo Seco Canyon. The rectangle at the upper right shows the part of San Gabriel Canyon studied earlier.

1 shows the area studied previously along the North Fork of San Gabriel River (Krumbein, 1940).

Practically the entire drainage basin of Arroyo Seco is shown on the Mt. Lowe (Los Angeles County) topographic map (1934). Figure 2 shows the extent of the basin and indicates all tributaries of Arroyo Seco within the mountain belt. The head of the main stream rises at Red Box Gap, at an elevation of about 4600 feet. The stream flows generally west and south for about 14 miles to the mouth of the canyon at the southern apex of the map where the elevation is 1150 feet. The average gradient is about 240 feet per mile. The canyon is steep-walled, with numerous rock exposures. Chaparral covers the gentler slopes, and the canyons themselves are lined with oaks and other trees. The numbers and letters on Figure 2 refer to sample localities, which will be described under appropriate heads.

Miller (1928; 1934) has described the geology of the San Gabriel Mountains. In general, the canyon cuts through a complex of crystal-

line rocks of possible pre-Cambrian and late Jurassic or early Cretaceous age. Several faults cut the region and locally control the trend of the main canyon. The mountains themselves are elevated fault blocks rather than folded belts. The present paper is not concerned directly with

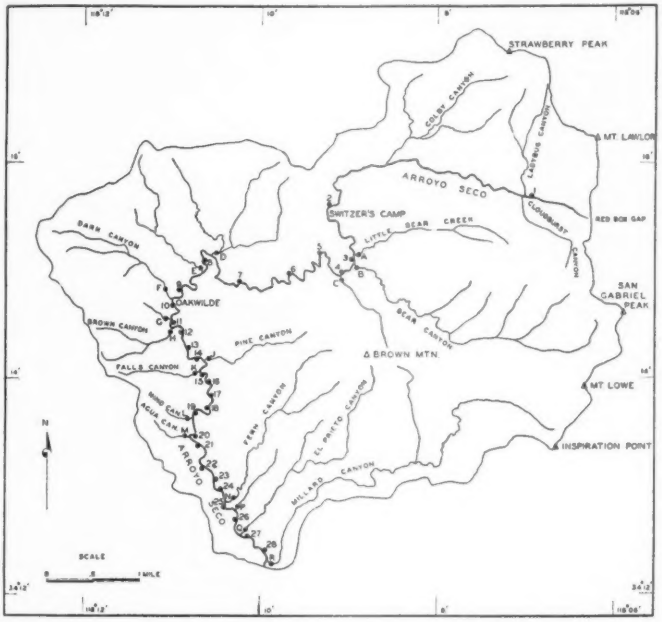

FIGURE 2.—*Arroyo Seco Canyon drainage basin*

the bedrock geology of the region, except to the extent that it controlled the kinds of debris. The most important rock in the drainage basin is the Lowe granodiorite of possible late Jurassic or early Cretaceous age, which underlies much of the eastern part of the drainage basin. Figure 3 shows the areal distribution of this granodiorite, taken from Miller's map (1934). The main stream cuts this igneous mass in the northeastern part of its course, but beyond its downstream point of emergence from the granodiorite it flows entirely through rocks distinctly different from the Lowe granodiorite. Moreover, between the unnamed tributary which enters the main stream below the northern bulge of the granodiorite (locality D, Fig. 2) and Millard Canyon (locality R, Fig. 2) no other tributary taps the rock. Thus the Lowe granodiorite

can be used to study the progressive rounding of pebbles during transport, with no danger of contamination in the lower half of the stream. Millard Canyon itself enters the main stream at the latter's point of emergence from the mountains and so does not change the situation.

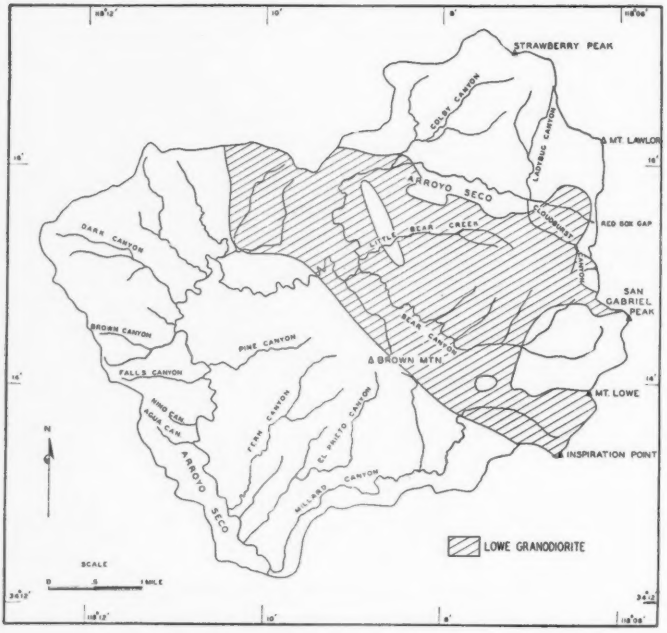

FIGURE 3.—*Distribution of Lowe granodiorite in Arroyo Seco*

FLOOD OF MARCH 2, 1938

GENERAL STATEMENT

The flood of March 2, 1938, equalled or exceeded any previous flood of record in Los Angeles County. The runoff peaks for the larger catchment areas equalled or exceeded any on record, and debris production was the greatest of record. Volumes of material eroded were estimated to vary from 12,000 to 80,000 cubic yards per square mile and averaged 51,000 cubic yards per square mile. This average is equivalent to the removal of more than a half-inch layer of debris from the entire area affected by the flood.

The Los Angeles Flood Control studied the engineering aspects of the flood (Burke, 1938), and through the kindness of Mr. Burke the follow-

ing details are given. Mr. D. M. Ilch of the U. S. Forest Service made available information on the sources of the flood deposits in Arroyo Seco, as well as estimates of the flood damage.

CONDITIONS BEFORE THE FLOOD

The catchment basin of Arroyo Seco supplies water for Pasadena. Conservation practices are followed, and although an attempt is made

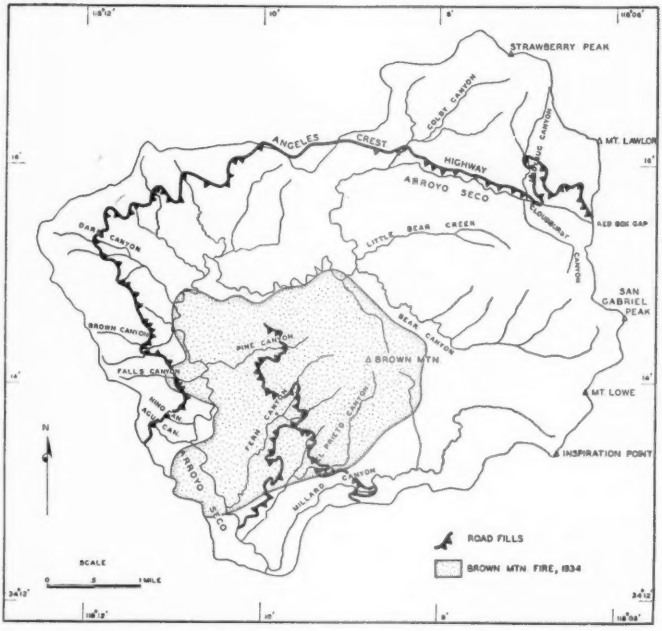

FIGURE 4.—*Sources of debris in Arroyo Seco flood of March 2, 1938*

to maintain a vegetation cover, in the past half century eight fires have swept parts of the basin. The most important recent fire, the Brown Mountain fire of 1934, burned an appreciable area in the center of the basin and made available a considerable amount of debris for movement during spring floods. Between 1934 and 1938 much silt had been carried down the tributaries into Arroyo Seco, and to a large extent this debris remained within the canyon area until the major flood. Within the last 10 years also, the Angeles Crest Highway was built through the mountainous area, involving numerous fills at steep canyon heads. During the 1938 flood this relatively loose material was avail-

able for transport by the flood waters. In addition, the steep mountain slopes were veneered with weathered debris which became unstable during the flood and caused numerous landslides. Fern Canyon, Pine Canyon, and Bear Canyon, as well as some parts of the main valley

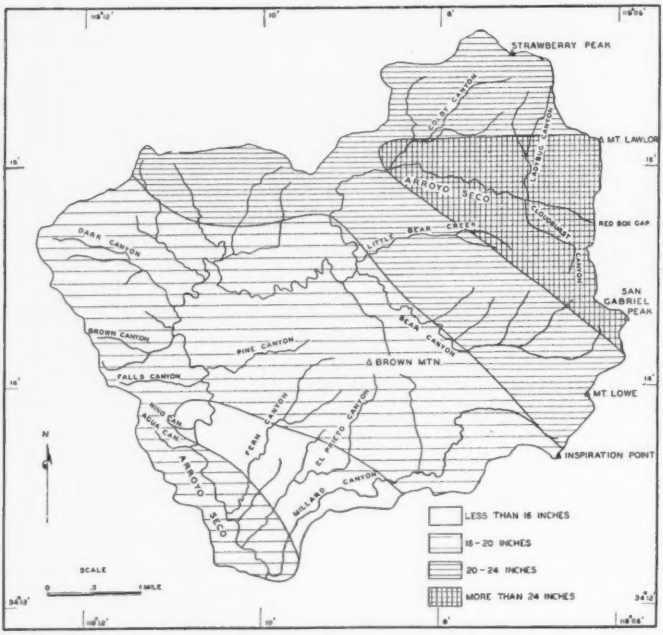

FIGURE 5.—*Generalized isohyetal map of Arroyo Seco*

itself, lie within or adjacent to the area burned by the Brown Mountain fire of 1934 (Fig. 4).

Before the flood the Arroyo Seco drainage basin had considerable unconsolidated material. Normal spring floods moved part of this along steep tributaries to the main canyon, but the lessened slope in the main valley prevented transportation to the alluvial fan. The flood of March 2, 1938, however, caused a major shifting of debris along the main canyon.

PROGRESS OF THE FLOOD

Rainfall data.—The Los Angeles Flood Control compiled rainfall maps for the storm. Figure 5 is a generalized isohyetal map of the entire storm, from February 26 to March 3, 1938. In the Arroyo Seco catchment area the precipitation varied from less than 16 inches near the mouth of the

canyon to more than 24 inches near the headwaters. During the first several days of the storm the rainfall was moderate, but on the evening of March 1 it increased abruptly. The moderate rainfall saturated the mantle rock, so that the more intense downfall resulted in excessive runoff. Generally speaking, the rainfall during the last period of the storm amounted to about 50 per cent of the total precipitation, which means rainfall intensities as high as 12 inches or more during a 24-hour period in some portions of the mountainous area.

The runoff associated with the flood varied during the storm. Previous to the intense downfall it is estimated at between 15 and 20 per cent. This left between 5 and 8 inches in the mountainous areas to saturate the mantle rock. During the second part of the storm the runoff increased to between 33 and 64 percent, the larger runoff associated with the larger drainage basins. In Arroyo Seco Canyon, a drainage area of 30.6 square miles, the runoff during the peak flood was 43 per cent. This excessive runoff resulted in very rapid accumulation of flood waters on March 2.

Stream discharge.—Estimates of maximum discharge along the canyon were made by engineers from high water marks. At Oakwilde the maximum was of the order of 6500 cubic feet per second. At the Ranger Station (near site 26, Fig. 2) it was about 9000 cubic feet per second, and at the entrance to Devils Gate Reservoir, just below the mouth of Millard Canyon, it was 11,400 cubic feet per second. The velocity of the flood water varied considerably along the canyon; maximum values in constrictions are believed to be of the order of 20 miles per hour.

Movement of debris.—Estimates of debris movement during any one flood are difficult to make, but fairly reliable figures are available for the Arroyo Seco drainage basin. About 1 mile downstream from the canyon mouth Devils Gate Reservoir served as a gauge on both the runoff and debris movement. Previous surveys of the reservoir showed that between 1920 and 1934 the reservoir received debris at the rate of 1920 cubic yards per square mile of watershed per year. The total estimated debris for the flood of March 2, 1938, was 70,000 cubic yards per square mile, which is about a 35-fold increase in the rate of erosion. During this one flood it is estimated that debris movement into or immediately upstream of Devils Gate Reservoir amounted to 1319 acre feet.

The maximum 24-hour inflow of water into the reservoir during the flood was 7420 acre feet. From this figure and the siltation of the reservoir Burke calculated that the volume of debris moved was 17.8 per cent of the maximum 24-hour inflow.

Flood damage.—Flood damage in Arroyo Seco Canyon was estimated at about $380,000, aside from the loss incurred in reduced reservoir capacity. This was distributed as follows: summer homes and camps, approximately $68,000; water diversion works, $28,000; road damage along Angeles Crest Highway, $140,000; and Arroyo Seco road and other minor roads, $144,000. More than 320,000 cubic yards of fill had to be replaced in embankments along 13 miles of Angeles Crest Highway after the flood.

DESCRIPTION OF FLOOD DEPOSITS

GENERAL CONSIDERATIONS

The flood deposits along Arroyo Seco Canyon were best developed in its lower course. Near the headwaters, especially upstream from Switzer's Camp (Fig. 2), numerous scours contained large boulders and local debris. Fresh bedrock was exposed in many parts of the stream channel. Downstream, especially below Oakwilde, the valley floor was covered with debris. Plate 1 shows the contrast between these two parts of the valley.

The Mt. Lowe topographic sheet (1934) shows a rather narrow flood plain in the lower third of Arroyo Seco Canyon traceable upstream from Oakwilde. Beyond this point the valley narrows markedly and enters what is locally known as "Marble Canyon" with precipitous walls and a sinuous course. Some of the tributaries in this gorge enter the main canyon over waterfalls. Where the flood plain appears aggradation becomes predominant as is indicated by the rarity of rock exposures in the channel floor downstream, except at narrows and sharp bends, where scour was locally severe. Upstream from the flood plain, however, bare spots on the canyon floor are not uncommon, and the deposits are generally more limited in volume and frequency.

In detail the flood deposits of the upper and lower canyon show a marked response to local conditions. Variations in valley width, and especially bends in the canyon, greatly affect deposition and scour. Where the valley widens a plain of sandy gravel may be found; at bends are boulder and log jams varying markedly in size and thickness. At tributaries along the canyon are alluvial cones, some of which cross the greater part of the valley floor.

LOG AND BOULDER JAMS

General statement.—Log jams in San Gabriel Canyon have been described previously (Krumbein, 1940), but their better preservation in Arroyo Seco and the more detailed study warrant this fuller consideration. In longitudinal section the jams are lenses or wedges thickening

FIGURE 1. EXPOSED BEDROCK IN STREAM CHANNEL NEAR SWITZER'S CAMP

FIGURE 2. FLOOD DEPOSITS 1.5 MILES BELOW OAKWILDE

CONTRAST OF SCOURED AND FILLED PORTIONS OF ARROYO SECO

TERMINUS OF BOULDER JAM NEAR SITE 5

(Photo by Lyman Huff) Taken from canyon wall downstream from jam. Note bedded material at left and gradual increase of boulder size downstream to terminus. The original front slope of jam is preserved in the central portion of the picture, but some post-flood erosion occurred at the right. The present stream flows from right to left.

and coarsening downstream. The upper surface may lie essentially horizontal or slope gently downstream. In some jams the thickened downstream portion may locally rise above the general level.

Near the upstream end the material is frequently relatively fine, moderately well bedded, and fairly well sorted. Downstream the deposit thickens and coarsens, and bedding becomes more prominent. Near the downstream terminus appear numerous large boulders, both on the surface and within the deposit; the sorting becomes noticeably poorer; and the bedding appears to give way to a heterogeneous arrangement. The jam terminates abruptly in a relatively steep slope, which drops to an area of scour immediately downstream of the deposit. This scour area contains lag boulders and cobbles and merges into the next flood deposit still farther downstream.

Plate 2 shows the terminus of a boulder jam near site 5 in Arroyo Seco. The photograph was taken from the canyon wall along the scour area below the jam, and the view is diagonally upstream. The marked increase in boulder size is shown, as well as the heterogeneity of the boulders. On the left is the finer relatively well bedded material upstream. Generally this ideal arrangement is complicated by several stages of reworking. One jam may partially overlap another, or one deposit may be partially destroyed and a later accumulation may be formed at approximately the same location.

Figure 1 of Plate 3, shows the bouldery texture near the downstream terminus of a jam deposit, and Figure 2 of Plate 3 shows the size variation along a jam which has been cleared parallel to the valley. This figure demonstrates that the boulders actually are concentrated near the downstream end and are not merely buried by the upstream portion. Plate 4 shows the openwork seam at the base of the flood deposits and the degree of imbrication often observed in the central portions of the deposits. (The pencil in the figure points downstream.)

A well-preserved log and boulder jam was found at site 25 in Arroyo Seco (Fig. 6). The valley turns abruptly and narrows downstream. A bridge had been built at the turn, and the old road ran along the east side of the canyon upstream from the bridge. When the site was studied, the main part of the jam was well preserved, although there were evidences of shifting positions of the stream since the flood. Upstream (Fig. 6) the deposits were relatively fine sandy gravel with a surface of lag cobbles. To the south and east boulders increased noticeably in number and size, and at the terminus of the jam the front slope dropped abruptly some 15 feet to an area of scour downstream. Here the valley was swept clean, and all the fill adjacent to the bridge founda-

tions had been removed. Fresh scours were visible on the exposed rocks along the valley wall, and except for large lag boulders this part of the valley was relatively free of debris. An idealized longitudinal section

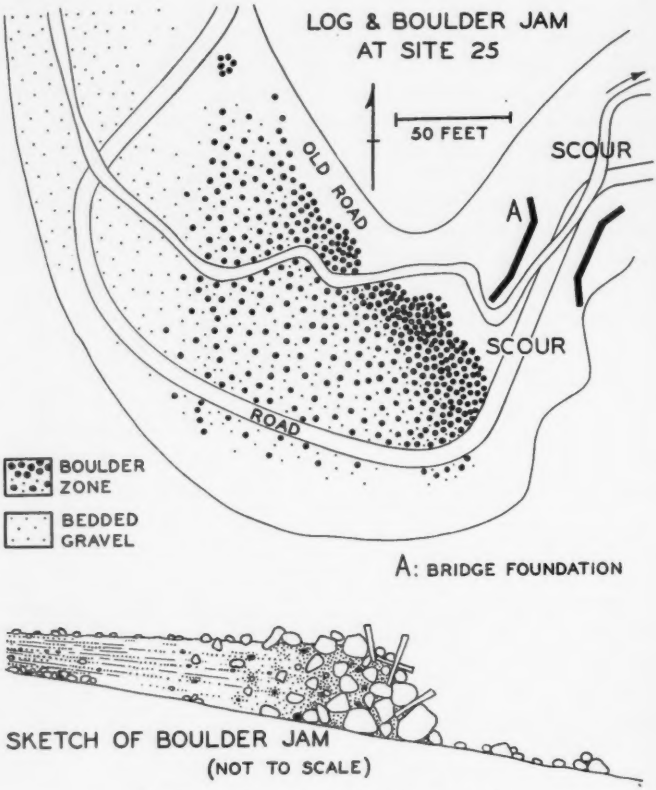

FIGURE 6.—*Map of boulder jam at site 25*
Lower figure is a sketch of the longitudinal section.

of the jam is shown in Figure 6. Numerous standing trees were associated with the deposit, but these are not shown.

Size variation of boulders.—The striking changes in size of the jam material is illustrated by parts of a small jam near site 21 in Arroyo Seco (Pl. 5). The photographs were taken approximately 35 feet apart, and the adjacent strips were mounted close together to bring out the

FIGURE 1. DOWNSTREAM END OF JAM NEAR SITE 8, SHOWING BOULDERY TEXTURE

FIGURE 2. LONGITUDINAL SECTION ALONG JAM CLEARED PARALLEL TO SIDE OF VALLEY
This photo shows that the boulders are actually concentrated at the downstream end, and not merely buried by the upstream portions.

FIGURE 1. OPENWORK SEAM NEAR BASE OF DEPOSIT, SITE 14

FIGURE 2. IMBRICATION OF PEBBLES IN CENTRAL PORTION OF DEPOSIT NEAR SITE 18
Pencil points downstream.

DETAILS OF JAM DEPOSITS

change in size. The downstream direction is downward in the plate. The largest boulder is about 4 feet in diameter.

The boulders in several jams were measured to estimate the magnitude of the size change. The material was too coarse for complete mechanical

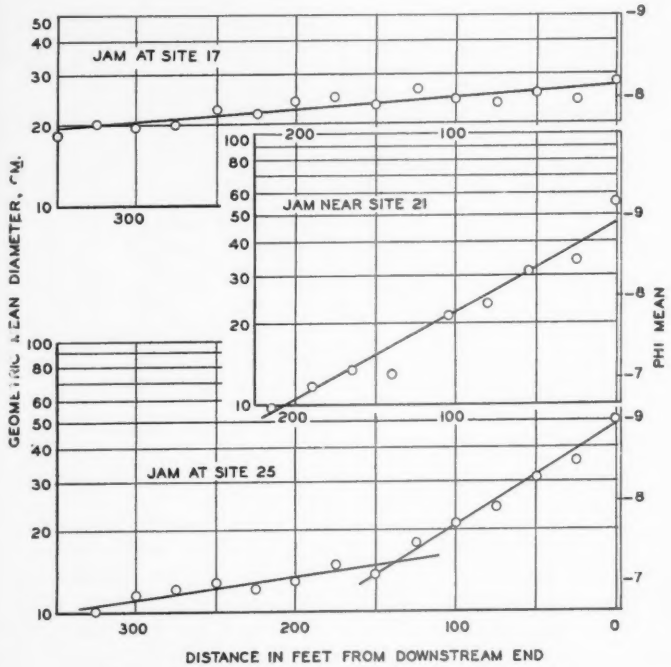

FIGURE 7.—*Size variation of boulders in jam deposits*

analyses, but for a first approximation sites were chosen at intervals of 25 feet along the jam, and the 20 largest boulders or cobbles at each site were measured. A knotted twine was held parallel to the intermediate diameter of each boulder, and this diameter was classified according to the Wentworth scale. The size classes ranged from 64 mm. (about 3 inches) to 2048 mm. (about 82 inches). The 20 values at each site were averaged to find the geometric mean boulder size (Fig. 7). In the jam near site 21 (Fig. 7, middle graph; *compare* Pl. 5), measured over a distance of 200 feet, the mean boulder size increased from 100 mm. to nearly 500 mm. In the jam at site 17 (Fig. 7, upper graph) the increase along 300 feet was from 200 mm. to 300 mm. At site 25 the slope

of the line (Fig. 7, lower graph) increases along the jam, suggesting a composite deposit.

Inasmuch as size is plotted on a logarithmic scale, the straight lines (Fig. 7) suggest that the downstream increase in boulder size is exponential. Such a change downstream must require special conditions of selective transportation, as the size gradient is reversed with respect to what may be expected along a stream. The slope of the straight line is a measure of the rate of size change and in the lower parts of the jams at sites 21 and 25 there is a doubling of the mean boulder size every 100 feet.

The decrease in large boulder size was not regular from one jam to another downstream; the largest boulders in the jams were just about as large near the mouth of the canyon as they were 10 miles or more upstream. Apparently the numerous tributaries furnished an adequate supply of material along the entire canyon so that no net decrease in large boulder size took place. When boulders of a particular rock type are considered, however, the mean size decreases downstream along the canyon. Thus boulders of Lowe granodiorite show a marked size variation.

Structure of jam deposits.—In the central, well-bedded portions of the log and boulder jams are usually three rather well-defined sedimentation units (Otto, 1938). The lowermost portion of the deposit is generally coarse gravel with cobbles, often displaying an openwork structure. Above this is medium-sized sandy gravel, usually fairly well bedded and imbricated. At the top is a thinner layer of coarse gravel and cobbles,—apparently a lag deposit. Occasionally the lowermost unit is medium sand (Pl. 4, fig. 2). In Plate 4 the three units are well developed, although the bedding of the central portion is not so prominent as in some instances.

In describing the jam deposits in San Gabriel Canyon, the writer concluded (1940):

"Some of the coarse lower material may consist of boulder bars formed behind fixed obstructions in the channel floor. In many cases, however, this lower material probably represents debris which dropped immediately upon initiation of a log jam. A considerable portion of finer material near the stream bed was entrapped with it, resulting in a poorly sorted deposit. This lower sedimentation unit may extend indefinitely upstream because of the slackwater effect produced by the temporary damming. Meanwhile, heavy deposition continued in the ponded areas near the log jam, the debris being supplied by the partially 'cleared' waters which continued to pour in from upstream. In this manner the intermediate well-bedded sedimentation unit accumulated. The upbuilding of the beds forced the stream to cut laterally around the jam unless the jam itself was partially undermined by the clear water pouring over its edge. In either case a channel lower than the surface

SIZE VARIATION ALONG JAM DEPOSIT

Photos taken about 35 feet apart along jam near site 21. (Downstream direction from top to bottom.) The largest boulder is about 4 feet in diameter. All photos to same scale.

FIGURE 1. DEBRIS CONE FROM SMALL TRIBUTARY NEAR SITE 17

FIGURE 2. ALLUVIAL PLAIN DEPOSIT NEAR SITE 9
Post-flood erosion has developed a low terrace in these deposits.

NONJAM DEPOSITS IN ARROYO SECO

of the deposit was soon developed. As the water collected into this new channel some of the topmost layers were swept downstream, leaving behind a lag deposit of cobbles and boulders, which is the topmost sedimentation unit."

This study supports the above conclusion, although some variations in the conditions of sedimentation may have occurred. The lower sedimentation unit was composed of poorly sorted gravel, of openwork seams, or of medium sand. The sand was found immediately upstream and downstream of a narrow gorge and may represent slackwater deposits formed during rising stages of the flood.

DEPOSITS AT TRIBUTARY MOUTHS

Deposition in the canyon was also associated with the mouths of many tributaries. Most of the tributaries have very steep gradients, and those entering from the west below Oakwilde contributed much debris, in part from the highway fills above. One prominent cone is at the mouth of Brown Canyon (site H, Fig. 2), another at the mouth of a minor tributary near site 17, leading directly down from the highway above (Pl. 6, fig. 1). Some tributaries supplied material from the highway fill, others carried large amounts of debris down from the burned-over area (Fig. 4). Fern and El Prieto Canyons, for example, had marked cones near their mouths composed of a rather heterogeneous mixture of fine and coarse material, with no evidence of sedimentation units such as those associated with the jams.

Immediately downstream from the tributaries the lithology of the main flood deposits usually reflected the contributions from the tributaries. In at least two instances the waters of the tributary and main stream evidently moved downstream side by side for an appreciable distance. The most marked occurrence of this sort was associated with Brown Canyon. This tributary contributed so much material to the main valley that the latter was literally choked; remnants of a large debris cone were plainly visible. This nonmixing of waters was discovered accidentally while pebble counts were being made. Along the main valley the pebble counts were expressed in terms of their content of Lowe granodiorite pebbles, which were used as an index of contamination. (The full discussion of pebble lithology is given later.) At site 11, 34 per cent was granodiorite pebbles, and at site 12 only 6 per cent. Additional samples from several local sites showed a marked and systematic variation in the content of granodiorite pebbles along this part of the valley. Figure 8 is a map of the field relations. Counts were made at the 29 sites indicated by numbers, which show the percentage of granodiorite pebbles at each point. The area studied was limited by the continuity of the deposits. At the downstream end the tempo-

rary road and shifting channels of the stream rendered it doubtful whether reliable counts could be had, and upstream, near the mouth of Brown Canyon, tractoring had destroyed the continuity of the deposits.

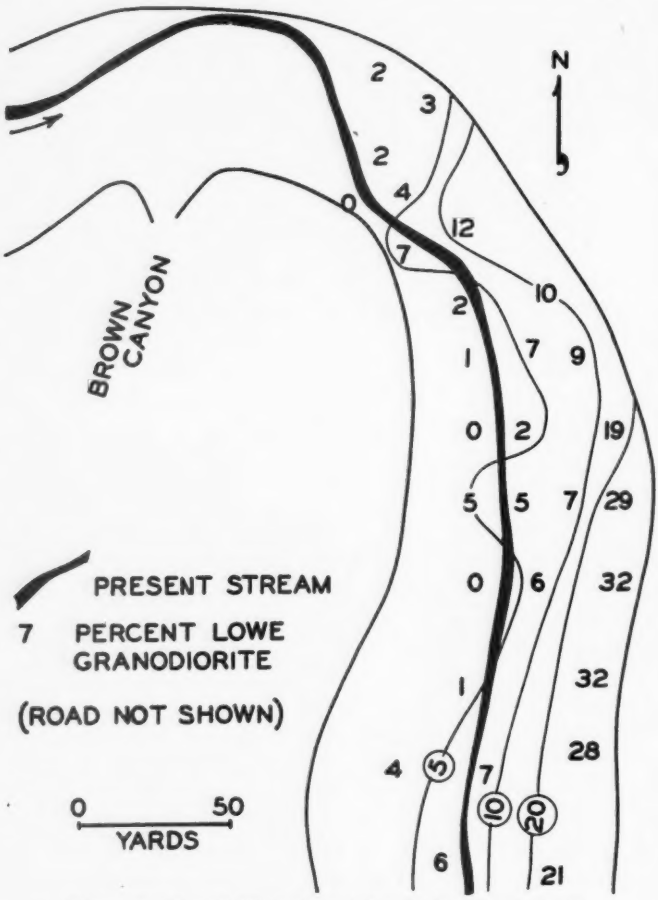

FIGURE 8.—*Map of lithologic variation in deposits at site 12*

Showing effects of nonmixing of tributary and main stream waters. The figures refer to percentage of Lowe granodiorite pebbles at each sampling site.

The rapid change in granodiorite pebble content is reflected by the isolines, which are geometric in interval. The linear pattern in Figure

8 reflects the parallel movement of tributary and main stream waters below Brown Canyon. Along the west side granodiorite pebbles were rare or absent, whereas along the east side they were abundant. The mixing of the waters was apparently confined to a narrow central zone. The geometric spacing of the isolines indicates the relative abruptness of the change from none to about 30 per cent of granodiorite pebbles.

In the northern part of the map the granodiorite pebbles are almost entirely lacking over the entire width of the valley for a limited distance, in definitely undisturbed deposits. Possibly the large debris cone from Brown Canyon may have effectively choked the main valley, forcing the main flood waters, already loaded, to the north wall of the canyon. This constriction in the valley may have increased the velocity of the main flood waters as in a conduit, with the result that transportation occurred through the conduit without significant deposition. As the valley widened around the bend, the flood waters spread out, mixed in part with the heavily charged waters from Brown Canyon, and began deposition.

Zones of delayed mixing probably occurred at several tributaries, inasmuch as a number of them contributed significant amounts of debris, especially from Angeles Crest Highway. Below Dark Canyon, where delayed mixing also took place the deposits were too discontinuous for detailed mapping. As with Brown Canyon, the granodiorite pebbles increased markedly from west to east.

DEPOSITS IN BROAD PARTS OF CANYON

In the lower portion of Arroyo Seco, where the flood plain was present, the relatively broad and straight stretches were covered with a more or less uniform plain of sandy gravel. One of the best developed of these alluvial plains was found at site 9 (Fig. 2). Just upstream of the site is a sharp bend in a narrows, whereupon the canyon becomes relatively broad and straight for about a quarter mile. Figure 2 of Plate 6 shows the deposits, now present as a terrace caused by post-flood erosion. At the downstream end of the stretch is another sharp bend and narrows, but no marked log or boulder jam. The main deposit apparently thins somewhat above the narrows, descending along a steeper gradient to a zone of relative scour with lag boulders which occurred in the narrows associated with the bend.

The deposits in the straight portion showed three sedimentation units as in the jam deposits, so that very likely similar conditions of deposition prevailed, although the ponding may not have been so abrupt as in true jams. Instead, the downstream narrows controlled much of the deposition, for it increased the depth of water to accommodate the discharge,

with the result that a backwater effect extended upstream, reducing the velocity and causing deposition. Meanwhile the rapidly moving partially cleared waters which entered the narrows produced a scour zone.

The three sedimentation units in this alluvial plain can apparently be related to early boulder bars or "pavement" over which the flood waters moved, and which now forms the bottom unit. The central bedded unit was formed during the bulk of sedimentation, and the topmost pebble and cobble lag was left behind by later waters which selectively reworked the top of the deposit.

The comparative simplicity of the picture at site 9 was elsewhere complicated by jam deposits which may have formed in conjunction with the alluvial plains in other parts of the canyon. Although the pronounced jams were often associated with bends, they sometimes occurred along a relatively straight stretch, where slight narrowing or other factors may have allowed a locking of logs or boulders. Therefore, the flood deposits presented a heterogeneous picture at first glance, which gradually resolved itself into the several elements as field work progressed. In some instances it was not possible to decide definitely which type of deposition predominated.

The canyon as a whole therefore presented a series of partially discontinuous deposits in its upper portion, succeeded by more continuous deposits downstream, with local regions of scour, and composed of partially distinct, partially overlapping deposits of jams, alluvial plains, and tributary-formed cones.

LITHOLOGY, TEXTURE, AND IMBRICATION OF FLOOD DEPOSITS

COLLECTION OF SAMPLES

To supplement the field data with more quantitative information, samples were collected for detailed laboratory study. Pebble counts were made at sites along the main valley and at every tributary mouth. Mechanical analyses were made along the lower portion of the canyon, where the deposits are well developed. Shape and roundness studies were made on pebbles downstream from Switzer's Camp (Fig. 2) to the canyon mouth. Fabric analyses were made of 11 samples, mainly in the lower part of the canyon.

Switzer's Camp was chosen as an arbitrary origin for purposes of graphing. It is in the central portion of the Lowe granodiorite area and is the highest point along the valley at which detailed analyses were made, with the exception of one oriented sample. All graphs given in later sections are referred to the origin at Switzer's Camp for convenience of comparison.

The samples were collected from the central, generally well bedded sedimentation unit at each sampling site. The samples for mechanical analysis were made by combining several samples through this unit at closely spaced intervals. Many of the samples were collected directly from jam deposits, but some of them were taken from cut banks in relatively straight parts of the valley, where the deposits were arranged in more regular alluvial plains. The following section summarizes the sampling data.

DESCRIPTION OF SAMPLES

Each item is keyed to indicate the type of sample taken. M = composite channel sample for mechanical analysis; L = sample for pebble count; S = shape analysis; R = roundness analysis; and O = sample of 100 oriented pebbles for fabric analysis.

Sample 1. Cut bank near center of valley, 2.75 miles upstream from Switzer's Camp. L, O.

Sample 2. Small bank along stream at Switzer's Camp. L, S, R.

Sample 2a. Talus sample (16-32 mm.) from steep ravine in Lowe granodiorite at Switzer's Camp. S, R.

Sample 3. Cut bank 0.95 mile downstream from Switzer's Camp. L, S, R.

Sample 4. Bank along stream, 0.3 mile downstream from sample 3. L, R.

Sample 5. Cut bank 0.45 mile downstream from sample 4. L, R, S.

Sample 6. Bank 0.8 mile downstream from sample 5. L, R, S.

Sample 7. Cut bank 1.0 mile downstream from sample 6. L, R, S.

Sample 8. Bank near east wall, 1.0 mile downstream from sample 7. L, S, R.

Sample 9. Bank in low terrace near west wall, 1.0 mile downstream from sample 8. M, L, R, S, O.

Sample 10. Cut bank in low terrace just below Oakwilde, 0.3 mile downstream from sample 9. M, L.

Sample 11. Cut bank in jam deposit, 0.3 mile downstream from sample 10. M, L, O.

Sample 12. Bank in low terrace, 0.25 mile downstream from sample 11. M, L, R, S.

Sample 13. Cut bank near east wall of valley, 0.25 mile downstream from sample 12. M, L, O.

Sample 14. Poor bank near center of valley flat, 0.25 mile downstream from sample 13. M, L, R.

Sample 15. Cut bank in jam deposit, 0.25 mile downstream from sample 14. M, L, O.

Sample 16. Bank in composite (?) jam, 0.2 mile downstream from sample 15. M, L, R, S.

Sample 17. Low bank in flood terrace, 0.25 mile downstream from sample 16. M, L, O.

Sample 18. Cut bank in jam near center of valley, 0.25 mile downstream from sample 17. M, L, R.

Sample 19. Cut bank in jam, 0.25 mile downstream from sample 18. M, L, O.

Sample 20. Bank in jam deposit, 0.25 mile downstream from sample 19. M, L, R, S.

Sample 21. Bank in low flood terrace, 0.2 mile downstream from sample 20. M, L, O.

Sample 22. Poor bank near center of valley, 0.3 mile downstream from sample 21. M, L.

Sample 23. Cut bank near east wall, 0.25 mile downstream from sample 22. M, L, O.

Sample 24. Bank in jam deposit, 0.2 mile downstream from sample 23. M, L, R, S.

TABLE 1.—*Size composition of Arroyo Seco flood deposits*

(Expressed as weight percentages)

Sample	Grade limits in mm.											
	128–64	64–32	32–16	16–8	8–4	4–2	2–1	1–1/2	1/2–1/4	1/4–1/8	1/8–1/16	–1/16
9	.....	7.6	14.1	14.4	13.0	15.2	12.2	10.5	7.5	3.7	1.1	0.4
10	.....	5.5	10.0	13.4	14.5	13.5	13.6	11.5	9.5	5.7	2.0	0.8
11	.....	14.2	19.5	17.0	9.4	9.3	8.2	8.3	7.5	4.2	1.7	0.7
12	.....	9.2	12.8	11.9	9.5	15.5	16.1	12.5	6.4	3.2	1.6	1.3
13	.....	12.0	18.0	10.1	7.8	11.0	12.5	12.6	9.1	4.3	1.7	0.9
14	.....	9.1	23.2	14.1	9.1	8.3	8.0	11.4	10.4	4.5	1.4	0.5
15	.....	7.0	11.9	12.8	11.6	13.4	15.0	13.6	8.2	3.9	1.6	1.0
16	.....	7.2	19.0	10.7	8.3	10.8	13.3	14.7	8.7	4.2	1.9	1.2
17	.....	11.6	14.7	12.1	13.5	15.9	13.1	10.1	5.0	2.2	1.0	0.8
18	.....	8.0	17.7	16.4	10.6	12.6	10.9	9.0	7.0	4.8	2.1	0.9
19	.....	12.4	20.9	14.6	9.0	10.3	11.4	11.0	5.7	2.5	1.2	1.0
20	.....	9.3	15.0	13.9	9.3	10.0	12.6	12.7	9.7	5.3	1.5	0.7
21	3.6	9.5	11.9	12.2	9.4	10.2	14.1	15.0	8.2	3.9	1.4	0.6
22	.....	17.5	17.0	15.6	11.1	9.0	5.5	6.3	8.5	5.9	2.4	1.2
23	.....	21.7	18.8	12.0	7.6	8.6	8.2	8.3	6.6	4.7	2.3	1.2
24	3.2	10.4	20.0	15.4	11.2	7.7	8.0	9.1	7.6	5.0	1.7	0.7
25	.....	12.1	15.9	16.7	12.5	9.0	8.7	9.9	8.7	4.7	1.5	0.3
26	.....	10.1	14.7	11.9	8.1	9.9	11.1	14.5	10.7	5.9	2.2	0.9
27	.....	11.6	11.9	13.7	10.8	12.8	13.3	12.1	8.0	3.5	1.4	0.9
28	1.0	10.6	14.5	15.2	12.4	14.0	10.9	8.3	5.5	4.3	2.4	0.9

Sample 25. Bank in jam deposit, 0.25 mile downstream from sample 24. M, L.

Sample 26. Cut bank in jam deposit, 0.25 mile downstream from sample 25. M, L, R, O.

Sample 27. Bank in jam deposit, 0.3 mile downstream from sample 26. M, L.

Sample 28. Cut bank near east wall, 0.25 mile downstream from sample 27. M, L, S, R, O.

Sample A. Little Bear Canyon. L.

Sample B. Bear Canyon. L.

Sample C. Unnamed tributary, 0.1 mile downstream from sample 4. L.

Sample D. Unnamed tributary, 0.25 mile upstream from site 8. L.

Sample E. Unnamed tributary, 0.2 mile downstream from site 8. L.

Sample F. Dark Canyon. L.

Sample G. Twin Canyon. L.

Sample H. Brown Canyon. L.

Sample J. Pine Canyon. L.

Sample K. Falls Canyon. L.

Sample L. Nino Canyon. L.

Sample M. Agua Canyon. L.

Sample N. Fern Canyon. L.
Sample P. Unnamed tributary, 0.1 mile upstream from site 26. L.
Sample Q. El Prieto Canyon. L.
Sample R. Millard Canyon. L.

TABLE 2.—*Statistical parameters of size data*

Sample	Phi mean M_ϕ	Phi standard deviation	Geometric mean diameter in mm. GM_ξ
9	−2.23	2.48	4.70
10	−1.41	2.37	2.66
11	−2.32	2.56	5.00
12	−1.70	2.42	3.25
13	−1.84	2.58	3.58
14	−2.04	2.57	4.11
15	−1.87	2.48	3.65
16	−1.63	2.54	3.10
17	−2.11	2.33	4.31
18	−1.93	2.46	3.81
19	−2.23	2.46	4.70
20	−1.69	2.54	3.22
21	−1.81	2.58	3.50
22	−2.27	2.74	4.81
23	−2.42	2.86	5.50
24	−2.38	2.66	5.21
25	−2.09	2.52	4.26
26	−1.63	2.66	3.10
27	−1.82	2.46	3.52
28	−2.05	2.50	4.14

MECHANICAL ANALYSIS

Twenty samples were collected for detailed mechanical analysis along the main stream from site 9 above Oakwilde to the canyon mouth, at site 28. The problem of choosing sampling sites was complicated by the apparent heterogeneity of the deposits and by the extreme size variations within the jam units. As in the earlier San Gabriel study, the sampling was confined to the central well-bedded portions of the flood deposits. The data thus obtained permitted a comparison of mean size and degree of sorting from jam to jam and along the interspersed alluvial plain deposits.

Each sample for size analysis weighed approximately 100 pounds. Material from the central sedimentation unit was collected from several closely spaced vertical channels through the unit to obtain a representative sample. The gravel, including all the coarse and fine material, was shoveled or scooped into canvas bags. These samples were analyzed into Wentworth grades.

The mechanical analysis data are given in Table 1. The particle size ranged from 128 mm. down to approximately $\frac{1}{32}$ mm. The amount of subsieve material (less than $\frac{1}{16}$ mm.) was generally so small that the range was not extended downward. Most samples have their maxi-

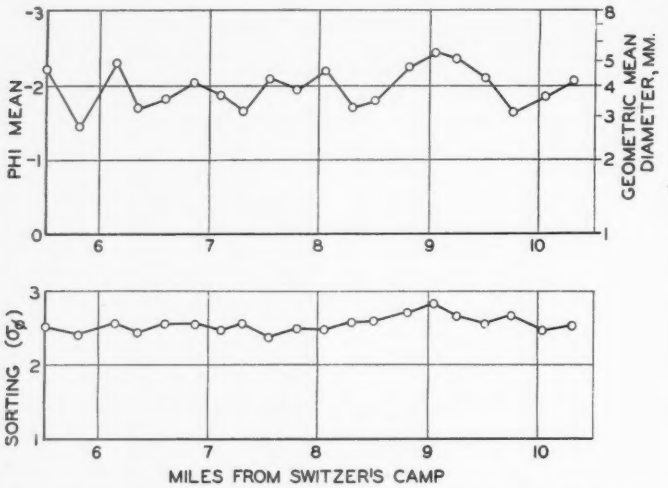

FIGURE 9.—*Variation in mean size and degree of sorting of flood deposits*
In central portions of jams and other flood deposits. The distance scale is referred to an upstream origin, for comparison with later graphs.

mum grade near the upper limit, and secondary maxima are not uncommon. Two statistical parameters were computed for each sample—the phi mean and the phi standard deviation (Krumbein and Pettijohn, 1938, p. 239-250). The phi mean was converted to its antilog, the geometric mean diameter in mm. These parameters are listed in Table 2.

Figure 9 is a graph of the variation of size and sorting along the valley. The broken line passing through the points does not strictly indicate the interpolated size between sampling sites, for each jam constitutes a unit, with considerable size variation from one end to the other. Interspersed among the jam units were more regular flood plain deposits in the straight and relatively broad parts of the valley. The samples were spaced approximately equally, so that they represent both kinds of deposits, depending upon which was present at a given sampling site. The present graph compares the central parts of succeeding deposits and as such shows a remarkable uniformity both in average size and degree of sorting from one deposit to the next. The absence of a marked

trend indicates that variations are nonsystematic and that the individual flood deposits represent essentially similar dynamic responses to transporting conditions all along the canyon. These data bear out a similar conclusion reached in San Gabriel Canyon.

LITHOLOGICAL ANALYSIS

In the size samples pebble counts averaged about 250 pebbles each. For convenience of comparison the counts were confined to the 16-32 mm. range. The pebbles were first classified into half a dozen categories according to rock type and were then grouped into classes which reflect to some extent the known areal geology of the region. The final classification adopted is as follows:

DOMINANTLY LIGHT-COLORED ROCKS:

(1) Pink granite, coarse to fine texture, locally pegmatitic, porphyritic, or aplitic. Dark mineral usually biotite or hornblende, irregularly scattered through rock. Quartz prominent. This group includes the Echo granite of Miller's report (1934).

(2) White granodiorite, moderately coarse textured, locally gneissic. Hornblende occurs in slightly elongated and somewhat irregular patches; in general, the dark minerals constitute less than 5 per cent of the rock. Small garnets fairly common in some facies. Quartz locally prominent. This is the Lowe granodiorite of Miller's report (1934).

(3) Gray granodiorite, coarse to fine texture, varying from quartz monzonites to quartz diorite. Dark mineral mainly hornblende; some biotite. This group of rocks is distinctly different from the white granodiorite in appearance, and little difficulty is encountered distinguishing them in the field.

(4) Fine-grained igneous rocks, typically felsites.

DOMINANTLY DARK-COLORED ROCKS:

(5) Dioritic and other coarse-grained basic rocks, medium to coarse texture, dark minerals predominant.

(6) Fine-grained dark rocks, including basalt, dike rocks, and greenstone.

(7) Foliated rocks, mainly schistose, with mica schist predominant. Includes well-developed gneisses and other banded rocks, usually dark.

Of these several categories the most outstanding group, and one of the most easily recognized, is the white Lowe granodiorite. The percentage of these pebbles decreased gradually downstream from site 9, and, since this decrease apparently indicated successive dilution of an original load, the pebble counts were extended to include the entire stretch of the valley from the Lowe granodiorite area (Fig. 3) to the canyon mouth. These samples were supplemented by pebble counts made at every tributary mouth along the same stretch, to round out the picture of lithological variation. A total of 45 pebble counts were made, and the data are listed in Table 3 on a percentage basis. The counts at the tributaries were made far enough upstream from their mouths so that possible backwater effects were eliminated. The data from the tributaries (samples A to R) show that, beyond the area of Lowe granodiorite proper, no tributaries penetrated into the gran-

TABLE 3.—*Lithological composition of 16–32 mm. pebbles, Arroyo Seco*
(Expressed as number percentages)

Sample	Light-colored rocks			Dark-colored rocks			
	Pink granite	White (Lowe) granodiorite	Gray granodiorite	Felsite	Diorite	Basalt, etc.	Foliated rocks
1	12	1	32	0	10	2	43
2	10	5	59	1	9	3	13
3	16	19	45	2	4	4	10
4	11	32	38	1	6	5	7
5	9	34	33	3	6	4	11
6	10	45	28	0	4	3	10
7	9	42	32	2	2	1	12
8	2	41	18	3	11	6	19
9	8	34	17	6	17	7	11
10	14	33	25	5	12	3	8
11	11	34	20	6	16	3	10
12	7	27	28	4	16	4	14
13	16	31	25	4	15	5	4
14	18	26	30	3	14	5	4
15	17	20	39	2	15	4	3
16	10	22	28	4	18	9	9
17	8	27	31	7	14	5	8
18	21	21	24	4	17	5	8
19	20	19	27	4	20	4	6
20	18	21	27	3	20	5	6
21	18	22	26	3	18	8	5
22	15	24	30	3	14	6	8
23	28	19	28	1	13	4	7
24	18	23	31	3	13	4	8
25	18	23	29	2	15	4	9
26	20	21	27	6	12	7	7
27	18	23	28	3	11	7	10
28	26	20	23	3	13	5	9
A	1	82	13	0	1	2	1
B	20	40	35	2	0	3	0
C	9	59	5	0	4	4	19
D	2	30	27	3	27	5	6
E	14	0	50	1	10	2	23
F	25	0	53	0	16	2	4
G	27	0	37	0	28	6	2
H	22	0	55	0	14	5	4
J	17	0	39	0	16	3	24
K	27	0	50	tr	20	3	tr
L	12	0	80	1	5	2	0
M	15	0	77	0	5	2	1
N	15	0	41	2	7	5	30
P	12	1	66	11	3	6	1
Q	10	0	47	0	6	4	33
R	30	17	31	2	8	1	11

odiorite except the unnamed tributary at site D and Millard Canyon at site R.

The data of Table 3 show that the quantity of Lowe granodiorite pebbles (third column) increases successively in the upstream samples,

reaches a maximum part way along the valley, and then decreases (Fig. 10). The rising part of the curve lies within the area of the Lowe granodiorite itself, and the maximum is reached at sample 6, near the western contact of the granodiorite. Within the Lowe granodiorite

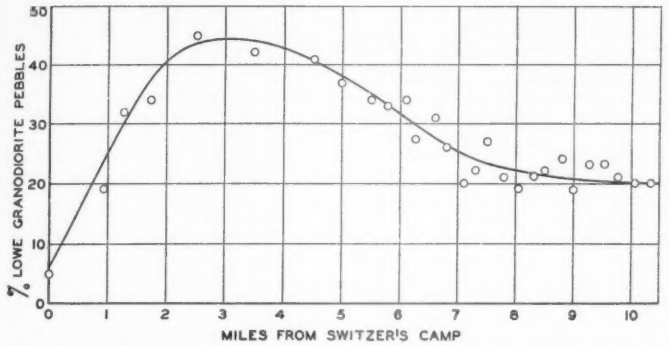

FIGURE 10.—*Variation in percentage of Lowe granodiorite pebbles in flood deposits along Arroyo Seco*

area practically all additions to the stream load are of that material, so that in passing through this section the load was continuously enriched by granodiorite pebbles. Beyond the downstream contact (with the exception of the tributary at site D) no more granodiorite was furnished by any tributary. The percentage of this rock thus diminished continuously as a function of the foreign contributions along the rest of the valley. The curve of decreasing granodiorite-content levels off downstream to an asymptote of about 20 per cent. This may indicate an approximate equilibrium between contributions from tributaries and the load coming down the main valley. Because the data are confined to the 16-32 mm. size range, however, this relation may be modified by a flow of material from one grade to another due to breakage and wear en route.

SAMPLING ERRORS OF SIZE AND LITHOLOGY ANALYSES

The samples for size analysis in the San Gabriel study averaged about 15 pounds each, and the question was raised whether these adequately represented the coarser sizes. In attacking this problem a series of samples was collected both in San Gabriel Canyon and Arroyo Seco, ranging from about 15 pounds to 300 pounds. Invariably the larger samples contained larger pebbles and cobbles, but in proportion to the total weight of the sample these had relatively little effect on the size parameters. In a 300-pound sample from Arroyo Seco, for instance,

was a cobble weighing 8.5 pounds, and yet, when the entire analysis was taken into account, the effect of this cobble on the median diameter was very slight. A 50-pound sample taken from the same site differed in its median diameter by less than 5 per cent. A comparative study

TABLE 4.—*Size variation of Lowe granodiorite boulders*
(Numbers represent arithmetic mean of 5 largest)

Distance downstream from Switzer's camp	Average size, meters
1.00	1.49
1.25	1.83
1.75	1.34
2.25	1.28
2.50	1.43
3.00	1.34
3.25	1.13
3.75	1.03
4.25	1.03
4.50	0.98
4.75	.76
5.00	.89
5.40	.79
6.00	.67
6.35	.61
7.10	.67
7.50	.55
7.90	.58
8.50	.52
8.70	.46
9.10	.55
9.60	.49
10.20	.58

of a set of six closely spaced samples in San Gabriel Canyon, of approximately the same weight as the first year's samples, showed a probable error of sampling of approximately 11 per cent. This is smaller than the observed variation from sample to sample along the canyon, but to reduce the sampling error it was concluded that samples twice as large would prove more satisfactory. The present size study in Arroyo Seco was based on 100-pound samples, for which the probable error of sampling is estimated to be less than 5 per cent.

The probable error of pebble counts was studied in a set of 24 closely spaced samples at site 9. The total probable error was found to be 10.2 per cent for sets of 100 pebbles, in terms of their Lowe granodiorite content. Assuming this same order of variability along the canyon, the samples of from 200 to 300 pebbles which were used for the data of Table 3 reduced this error to about 6 per cent.

SIZE VARIATION OF LOWE GRANODIORITE BOULDERS ALONG THE CANYON

A special supplementary study was made of the Lowe granodiorite boulders themselves. Boulders of all compositions in the jams are about the same size all along the canyon. It was anticipated, however, that

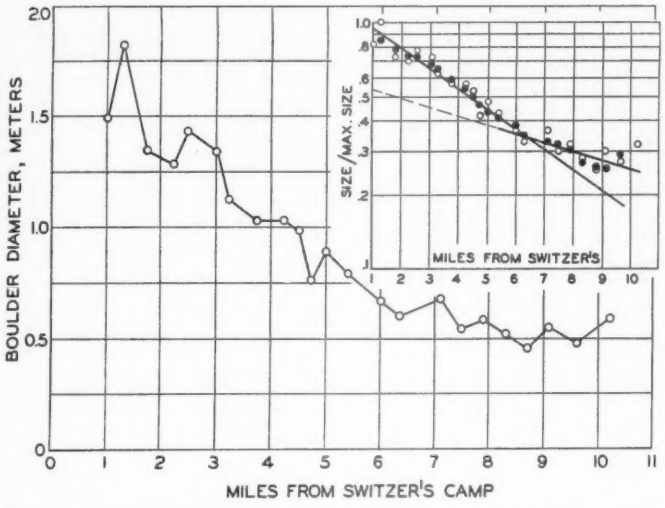

FIGURE 11.—*Variation in size of Lowe granodiorite boulders along Arroyo Seco*

Inset shows same data on semi-logarithmic plot. The black dots are moving averages based on successive sets of three observations.

if measurements were made of Lowe granodiorite boulders themselves there should be a marked decrease in size down the valley, due in part at least to continued breakage and abrasion.

To obtain pertinent information Lowe granodiorite boulders were measured at intervals along the canyon. In the upper reaches the boulders were very numerous, but downstream they became comparatively scarce. For comparison, therefore, a number of boulders were measured at each site, and the average of the five largest was taken as an index. Table 4 lists the data, and Figure 11 shows the results. The data scatter rather widely, but a definite trend is discernible. The exponential trend of the data is brought out by the inset figure, which shows boulder size plotted on a logarithmic scale. The solid circles represent a moving average based on successive sets of three observations. Two straight lines are drawn through the averaged data; these suggest

a decrease in the rate of change of boulder size about 6 miles downstream from Switzer's Camp (and about 1 mile downstream of the point where the flood plain appears in Arroyo Seco). If the break in the curve does coincide approximately with the transition region over which, on the average, scour conditions are replaced by depositional

TABLE 5.—*Sphericity data of Lowe granodiorite pebbles, 16–32 mm. class*
(Frequency expressed as number percentages)

Sample	.40–.49	.50–.59	.60–.69	.70–.79	.80–.89	.90+	Mean sphericity	Standard deviation
2	2.2	11.1	53.3	26.7	4.5	2.2	0.68	0.088
3	2.6	13.1	29.0	47.4	7.9	.....	.70	.091
5	2.9	14.3	40.0	34.3	7.1	1.4	.68	.095
6	1.6	23.8	42.9	25.4	6.3	.....	.68	.086
7	.....	12.2	34.7	40.9	12.2	.....	.70	.086
8	1.3	6.4	38.5	42.3	11.5	.....	.71	.082
9	.....	10.0	36.7	41.1	11.1	1.1	.71	.086
12	.....	14.0	28.1	42.1	15.8	.....	.71	.092
14	2.4	3.6	36.1	39.8	18.1	.....	.72	.089
16	2.4	16.6	34.5	40.5	4.8	1.2	.68	.093
20	.....	5.8	37.9	46.0	9.2	1.1	.71	.079
24	2.2	12.2	41.2	32.2	12.2	.....	.69	.099
28	.....	17.8	32.2	43.6	4.8	1.6	.69	.092

conditions, the change of slope in the graph may have a physical significance. Scour conditions imply more vigorous wear and breakage, whereas depositional conditions imply relatively more selective transportation. Thus the difference in slope of the two lines may offer a first approximation to the influence of the breakage factor during transport.

Strictly speaking, this analysis of boulder size goes beyond the effects of a single flood; probably any given boulder moves only a relatively short distance during a single flood, so that the observed size variation is the cumulative effect of a number of movements. The size variation is a composite of the effects of selection during transport, breakage, and abrasion en route, and perhaps of scour by finer material while the boulders are at rest (Maxson, 1940). That one of these factors predominates in given parts of the canyon is likely, and with adequate data it should be possible to isolate some of the variables.

SHAPE AND ROUNDNESS ANALYSIS

San Gabriel Canyon carried a load which lithologically had much greater uniformity than did Arroyo Seco where numerous tributaries contributed a large part of the load. Observation as well as theory indicate

large variations in shape and roundness along Arroyo Seco, even within a given size range. Nevertheless, the Lowe granodiorite pebbles offered a means for studying both shape and roundness as functions of distance along the stream, because any pebbles in the lower part of the canyon

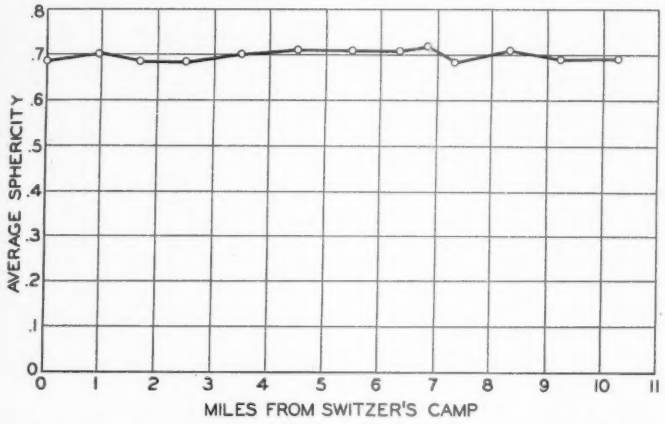

FIGURE 12.—*Variation in pebble shape (sphericity) along Arroyo Seco*
Lowe granodiorite, 16-32 mm. class.

could have reached that point only by traveling along the main valley. The situation is thus similar to Wentworth's Russell Fork study (1922).

Some of the samples collected for mechanical analysis contained sufficient pebbles for shape and roundness study. These were supplemented by samples collected upstream as far as Switzer's Camp to bring out the essential features of the curves. The shape (sphericity) was measured by a graphic intercept method based on axial ratios of the pebbles. This method is more rapid than earlier methods and gives results in substantial agreement. Roundness was estimated by a rapid visual method from a standard set of pebble images. Both methods are described in detail elsewhere (Krumbein, 1941a).

The sphericity data are listed in Table 5, which also includes the arithmetic mean sphericity of the samples. Figure 12 shows the variation in mean shape along the valley. The remarkably uniform values from point to point indicate a small nonsystematic variation, reflecting uniform conditions of shape selection along the valley, and thus indicating essentially similar dynamic conditions of deposition. The same feature was shown, but less clearly, in San Gabriel Canyon.

Figure 13 also brings out the absence of any marked change in pebble sphericity over the distance studied. The ratio b/a is the numerical value obtained by dividing the length of the intermediate pebble axis by the longest pebble axis. The ratio c/b is the ratio of the short axis to the intermediate axis. The curves represent lines of equal sphericity super-

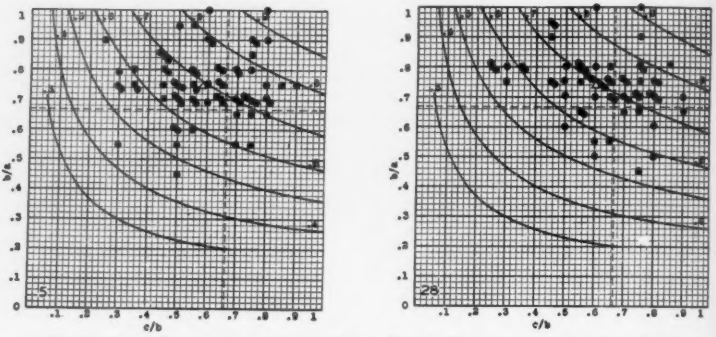

FIGURE 13.—*Comparison of sphericity distributions of samples 5 and 28*
See text for details of plotting.

imposed on the axial ratios (Krumbein, 1941a). The dashed lines are Zingg's shape classes (Zingg, 1935; Krumbein and Pettijohn, 1938, p. 288). Comparison of the two figures shows that there is no marked change in sphericity between the two sets. In both instances the majority of the pebbles have sphericity values between 0.6 and 0.8. The centers of gravity of both samples (shown by the triangles) lie at almost exactly the same positions. The degree of scatter of the data is also roughly the same.

The roundness data offered several difficulties, especially in the upper part of the canyon. Within the area of Lowe granodiorite new material was always available to the stream, so that throughout that stretch angular fragments were continuously added to a load of already partially rounded granodiorite pebbles. Once the stream leaves the granodiorite area no more granodiorite fragments are contributed,[1] so that the average roundness tends to become more uniform. The result is an inevitable scatter of the data among the upstream samples. To reduce this scatter, an average was taken of samples 2, 2a, and 3, which represent the first three samples in the granodiorite area. Sample 2a came

[1] An exception is the unnamed tributary at site D, but even here the pebbles have traveled a fair distance from the outcrops before they enter the main valley.

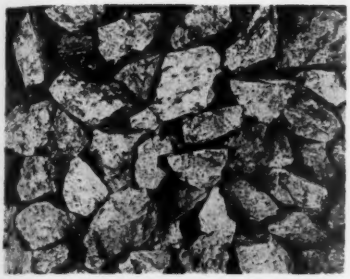

FIGURE 1. FRAGMENTS FROM TALUS AT SWITZER'S CAMP

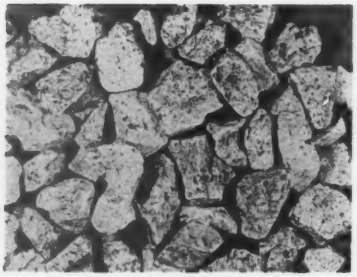

FIGURE 2. PEBBLES FROM CUT BANK AT SWITZER'S CAMP

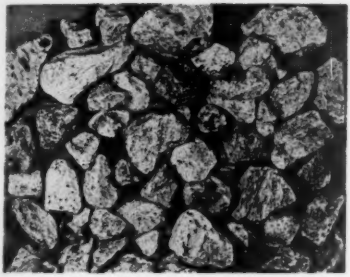

FIGURE 3. SAMPLE 7, 3.5 MILES DOWNSTREAM FROM SWITZER'S CAMP

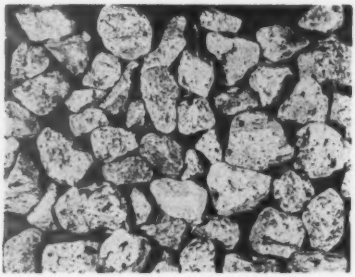

FIGURE 4. SAMPLE 8, 4.5 MILES DOWNSTREAM FROM SWITZER'S CAMP

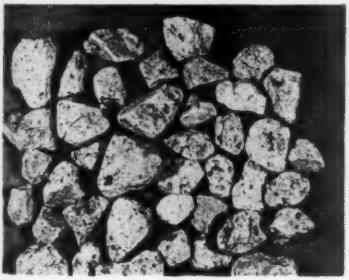

FIGURE 5. SAMPLE 22, 8.8 MILES DOWNSTREAM FROM SWITZER'S CAMP

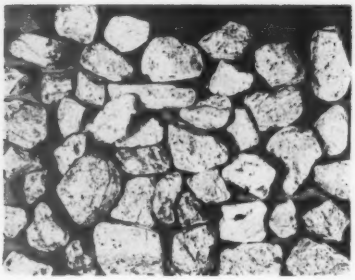

FIGURE 6. SAMPLE 27, 10.0 MILES DOWNSTREAM FROM SWITZER'S CAMP

ROUNDNESS VARIATION IN LOWE GRANODIORITE PEBBLES, 16–32 MM. CLASS

from a talus slope along the canyon and gives an indication of the initial roundness of the fragments.

Plate 7 shows the roundness variation in six samples. The total change is not very marked, but differences among the first few samples are greater

TABLE 6.—*Roundness data of Lowe granodiorite pebbles, 16–32 mm. class*

(Frequency expressed as number percentages)

Sample	Midpoints of roundness classes						Mean roundness
	.1	.2	.3	.4	.5	.6	
2	45.2	22.7	22.7	9.4	.....	.....	0.20
2a	98.5	1.5	.....	.....	.....	.....	.10
3	7.9	63.1	29.0	.....	.....	.....	.23
Composite*	57.5	24.2	15.0	3.3	.....	.....	.16
4	35.8	43.4	18.9	1.9	.....	.....	.19
5	17.6	50.0	23.0	8.1	1.3	.....	.23
6	5.8	55.0	29.0	7.3	2.9	.....	.25
7	.....	28.3	41.5	26.4	3.8	.....	.31
8	.....	8.3	50.4	36.3	5.0	.....	.34
9	.....	6.0	56.0	36.0	2.0	.....	.34
12	.....	2.2	51.1	44.5	2.2	.....	.35
14	.....	12.2	48.2	30.9	6.2	2.5	.34
16	.....	.....	57.1	37.2	5.7	.....	.35
18	.....	7.0	27.9	57.0	8.1	.....	.36
20	.....	2.6	41.0	48.6	7.8	.....	.36
24	.....	.....	57.0	41.4	1.7	.....	.35
26	.....	.....	42.1	50.0	7.9	.....	.37
28	.....	2.8	28.7	51.4	17.1	.....	.38

* Average of samples 2, 2a, 3.

than among the later ones, indicating more rapid rounding during the early stages of the process.

The roundness data are listed in Table 6, and the corresponding graph of arithmetic mean roundness is shown in Table 14. The data show a definite trend, leveling off toward an asymptote of about 0.40 downstream. To bring out this asymptotic value more clearly, several additional samples were measured near the lower end of the series. The inset of Figure 14 shows the roundness data plotted on semilogarithmic paper, assuming an asymptote of 0.40. The ordinates represent the difference between the asymptote P_0 and the value of the roundness P, divided by the asymptotic value. The straight line indicates the fair agreement of the data with an exponential trend, as discussed in the San Gabriel paper (Krumbein, 1940). The coefficient of rounding in the present instance is 0.23 as compared to 0.50 in San Gabriel.

As in the San Gabriel deposits, the roundness of the pebbles is a function of the distance transported. The smaller coefficient of rounding may be the result of one or more factors. The rocks involved may have had different resistances to wear; the velocity of the flood waters may

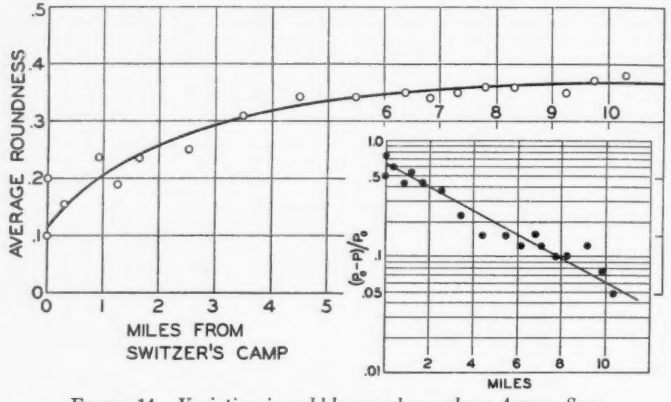

FIGURE 14.—*Variation in pebble roundness along Arroyo Seco*
Lowe granodiorite, 16-32 mm. class. Inset shows same data on semi-logarithmic plot.

have been different; and the Arroyo Seco pebbles were from a slightly smaller size range than the San Gabriel pebbles. The absolute magnitude of the coefficients is perhaps not so significant as their relative values. The ratio $0.50/0.23 = 2.2$ means that on the average the San Gabriel pebbles were rounded twice as rapidly as those in Arroyo Seco.

Neither the San Gabriel nor the Arroyo Seco data show the incipient steeper portion of the roundness curve found experimentally (Krumbein, 1941b). Field conditions may obscure this feature, although it was present in Wentworth's Russell Fork data, discussed in the abovementioned paper. Possibly, in the distances covered in Arroyo Seco and San Gabriel Canyon, the accelerated rate of rounding during the flood may extend the initial steep portion over a greater distance than under less rigorous conditions.

IMBRICATION OF PEBBLES IN FLOOD DEPOSITS

During the San Gabriel study, evidence was found of reverse imbrication in the sample collected at the highest upstream position. The general site was revisited during 1940, and several additional oriented samples were collected. Upstream from the original site a very large boulder set up a "shadow zone" of reversed imbrication on its down-

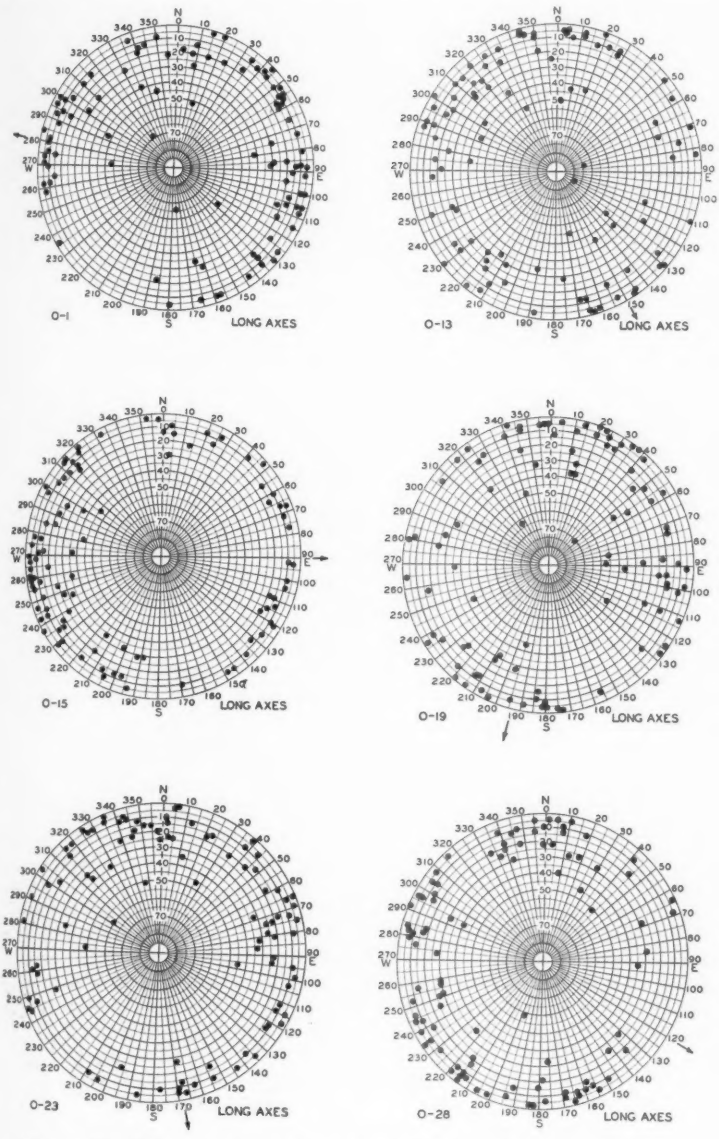

FIGURE 15.—*Polar co-ordinate charts of pebble orientation in Arroyo Seco*

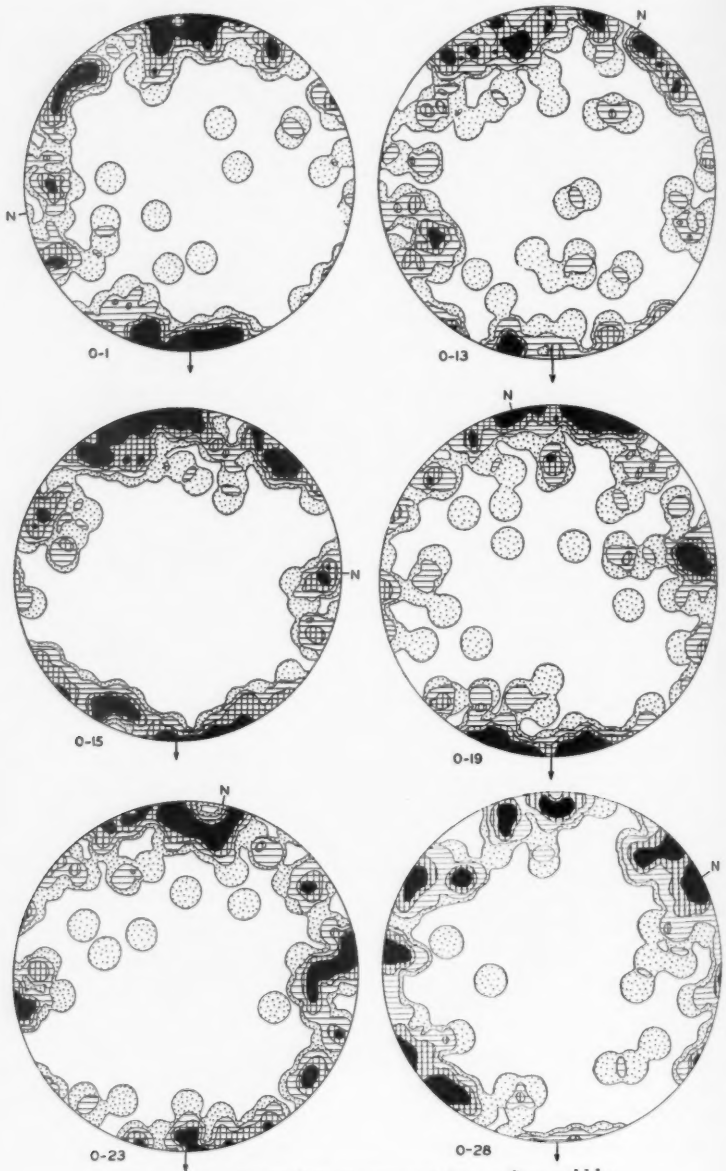

FIGURE 16.—*Petrofabric diagrams of Arroyo Seco pebbles*

stream side, extending nearly 150 feet and tapering off. Apparently sufficient disturbances were set up in the lee of the boulder to generate reverse currents or eddies, in the zone of which the deposited materials were imbricated by a reverse current. Other similar large boulders in

TABLE 7.—*Dip direction and average roundness of oriented pebbles (Arroyo Seco)*

Sample	Mean azimuth	Azimuthal standard deviation	Mean roundness
0– 1	106.2°	43.8°	0.28
0– 9	3.6°	46.4°	.32
0–11	51.0°	46.4°	.31
0–13	143.0°	46.0°	.23
0–15	79.2°	43.2°	.19
0–17	136.6°	46.2°	.24
0–19	21.0°	47.4°	.25
0–21	119.8°	51.0°	.32
0–23	155.4°	49.4°	.34
0–26	179.2°	48.2°	.32
0–28	168.2°	47.8°	.32

the canyon indicated that the magnitude of the "shadow zone" was a function of the boulder size.

In Arroyo Seco, by means of an orienting frame (Krumbein, 1939), one oriented pebble sample was collected in the steep upper part of the canyon (site 1, Fig. 2), and 10 samples were collected in the lower part of the valley, from sites 9 to 28. In every instance a normal upstream imbrication was found; Figure 15 shows the data of six typical samples. In Figure 15 any given point has the co-ordinates, dip direction (measured along a circumference), and angle of dip (measured inward along a radius). Figure 16 shows the corresponding petrofabric diagrams, all oriented so that the direction of valley trend is vertically downward. The diagrams show normal imbrication, including sample 0-1, which was collected at a point where the stream gradient was about 500 feet to the mile. In each instance the symmetry is monoclinic, with the *a*-fabric axis more or less parallel to the valley trend. Considering the numerous swirls, eddies, and reverse currents that may be present in floods, the general agreement of the preferred orientation to valley trend is significant. Apparently the mean azimuth may be safely considered as a statistical index of direction of stream flow.

In the San Gabriel study the azimuthal standard deviation of the pebbles about the mean preferred orientation varied as a function of

distance along the valley, but the present data do not show any systematic variation downstream. It was also noted in the earlier study that the azimuthal standard deviation was larger for rounded pebbles

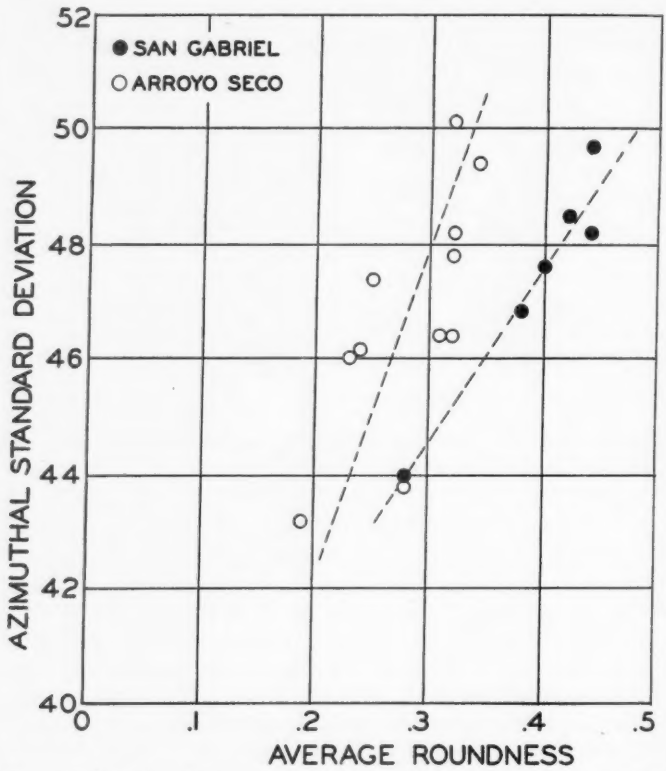

FIGURE 17.—*Scatter diagram of average roundness and azimuthal standard deviation of oriented pebbles*
Arroyo Seco and San Gabriel Canyon.

than for angular pebbles. The Arroyo Seco data do not clearly support this observation. Table 7 lists the azimuthal standard deviation of the 11 Arroyo Seco samples, as well as the average roundness of the oriented pebbles. The pebbles in these samples include all rock types, so that the average roundness does not agree in general with the average roundness of the Lowe granodiorite pebbles given in Table 6. Figure

17 is a graph of azimuthal standard deviation plotted against mean roundness. The open circles represent the Arroyo Seco samples, and the solid circles represent the San Gabriel samples (Krumbein, 1940, Tables 6, 7). The Arroyo Seco data scatter rather widely, whereas the San Gabriel data show a fair linear trend. In view of this unsatisfactory evidence, generalizations regarding the spread of pebble axes as functions of distance or of average roundness are deferred.

INTERPRETATION OF FIELD AND LABORATORY DATA

GENERAL CONSIDERATIONS

Some interpretations have been given in the descriptive portion of the paper, but an integration of the several lines of evidence should be made. Three general problems deserve discussion: (1) the mechanics of transportation of flood materials; (2) the physical factors which control their deposition; and (3) the interplay of wear and selective transportation during movement. These features must be discussed almost entirely from post-flood evidences as direct observational data on floods in steep canyons are very meager. Engineers are naturally concerned most with damage to business and residential districts, which confines their work during floods to regions beyond the canyon mouths. Stream gauges and other equipment along the canyons are usually destroyed or ineffective during large floods, so that many basic data are never obtained. Estimates of discharge based on high water marks are at best uncertain.

The geological approach offers the best solution until more quantitative observational data are available. Gradational processes are reflected in the characteristics of the deposits they form, and hence a study of the deposits should shed light on the mechanical processes involved.

Troxell and Peterson (1937), studied the effects of bends and constrictions on deposition in Pickens Canyon after the flood of January 1, 1934. Their conclusions are fully supported by the present study, insofar as similar conditions existed in Arroyo Seco. Pickens Canyon, near La Canada, California, lies some 5 miles west of the mouth of Arroyo Seco Canyon. It is much smaller with a canyon section about 2 miles long. Jam deposits are not specifically mentioned, but the mechanics of transportation and deposition along the canyon are presented. The following sections attempt to apply this work to Arroyo Seco and to extend it to include the effects on the character of the sediments.

MECHANICS OF MOVEMENT OF FLOOD DEBRIS

The movement of sediment by running water has been studied by numerous workers, and from these investigations a number of universal principles have emerged. Several generalizations are used here as a basis for a qualitative statement of the problem. The ability of a stream to transport material may be expressed as its capacity (the quantity it is capable of carrying per unit time) or as its competence (the sizes it can move). In general the competence increases as the capacity increases, but the relation is not simple. Gilbert (1914) showed that capacity increases with the slope and with the discharge, as well as with the fineness of the debris.

Bed load is carried along the bottom of a stream, and suspended load at some distance above the bottom. Saltation occurs in an intermediate zone. Researches by Lane and Kalinske (1939) show that the distinction between bed load and suspended load is partly a function of the turbulence intensity near the bed, which controls the sizes, shapes, and densities of particles that may be lifted from the bed. The continued movement of material in suspension depends upon a concentration gradient of solid material in the vertical section (O'Brien, 1933), so that turbulent diffusion may carry more material upward than is carried downward by the same process. This excess upward diffusion is counterbalanced by the settling velocities of the particles themselves, and the interplay of these two factors establishes an equilibrium condition of suspension transport.

The movement of bed load is a function of bottom velocity, and recent work (Shields, 1936; Rubey, 1937) has shown a relationship between the thickness of the laminar sublayer in the flowing water and the sizes and amounts of material moved. Bed load movement can also be expressed in terms of discharge or of average velocity, which are known to be related to transportation, but perhaps not so fundamentally as is the bottom velocity. Nevertheless, any factor which changes the average velocity of the water will also change the capacity.

In floods the conditions of stream flow are much exaggerated due to the high discharge. A stream in flood may have a discharge a thousand times as great as the same stream at mean low water; in large floods this ratio may be even higher. Although conditions are thus much more violent during floods, there seems to be no valid reason why the same underlying physical laws of sediment behavior should not hold. Perhaps the distinction between bed and suspended load loses some of its sharpness. The steep gradients in mountain canyons and the high degree of turbulence generated by irregular surfaces throw into temporary suspension relatively large material. Equilibrium con-

ditions, in which the loads become more or less distinct, may never be reached, so that the normal downward increase of particle size in a vertical section is confused by the intermingling of sand and pebbles at some distance above the bed. That such heterogeneity may have been present is indicated in part by the intimate mixture of coarse and fine material within the central sedimentation unit of both Arroyo Seco and San Gabriel canyons. Practically all size distributions are polymodal, suggesting a mixture of bed, saltation, and suspended load. Many exposures contain pebbles completely surrounded by sand, which renders it unlikely that the material represents two episodes of deposition, as an openwork gravel with the later infiltration of sand. Observations by Chawner (1934; 1935) also bear on this point. Splashings from the La Crescenta flood collected from roof drains showed on mechanical analysis a rather poorly sorted material with an appreciable content of pebbles as coarse as 8 mm. diameter. Apparently these splashings represent suspended material. There is little evidence of the relative importance of saltation and suspension in floods, but the high degree of turbulence in flood water suggests that if particles are freed from the bottom, turbulence intensity may be great enough to transfer them to the suspended load. The saltation zone in streams is usually rather shallow, whereas the main sedimentation unit in Arroyo Seco is locally several feet thick.

Such lines of evidence suggest that for some depth above the bottom the flood waters of Arroyo Seco carried a fairly concentrated load ranging from sand to moderate-sized pebbles. Toward the base of the zone cobbles and larger fragments rolled or slid along, behaving as bed material. The evidence points to simultaneous deposition from this concentrated zone, which would require a relatively abrupt velocity decrease to drop medium as well as coarse grains. No doubt the cessation of movement of coarse particles in such a load results in the trapping of some finer material, which may add to the content of sand and silt in the final deposit.

CONDITIONS CONTROLLING DEPOSITION

Figure 18 shows the plans and side elevations of two canyon sections; the upper has parallel sides and a uniform slope; and the lower has the same slope but contains a constriction. If water flows uniformly along the upper section, its depth remains constant, and the slope of its upper surface parallels the slope of the canyon bottom. Here sediment is transported at a given rate along the entire stretch. In the lower canyon the constriction completely changes the conditions. To maintain a given discharge the water in the constricted zone must move

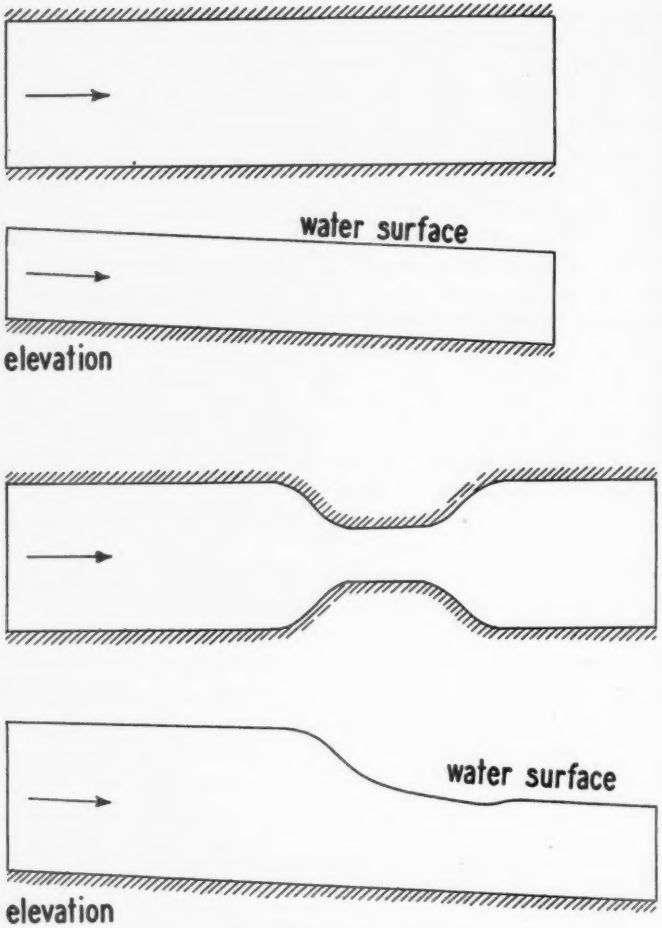

FIGURE 18.—*Plan and elevation of uniform and constricted channels*
Showing water profiles in steady flow.

more rapidly. This can be accomplished only as the slope of the water surface steepens, inasmuch as the slope of the canyon is fixed. The result is that the water deepens appreciably at the upstream end of the constriction and thus backs up the water in the broader part of the canyon. The depth of water in the broad part decreases progressively upstream at a rate too slight to be shown in the figure, but

over any considerable stretch average velocity changes. The net effect is that velocity decreases downstream until the narrows is reached, whereupon it increases rather markedly.

This analysis applies to mild slopes, in which the normal depth is greater than the critical depth (Rouse, 1938, p. 290 ff.). It also applies to steep slopes under the same conditions of depth. If the approaching flow is supercritical a hydraulic jump usually forms on the upstream side of the constriction. The jump may migrate upstream and thus develop the situation shown in Figure 18, with the jump left of the diagram. Whichever situation holds, it seems likely from the geological evidence that deeper water and lower velocities were developed upstream from the constrictions, thereby causing deposition, whereas within the constriction the velocity increased again and permitted scour.

A bend in the canyon associated with the constriction causes additional loss of energy. This effect may be slightly offset by an increase in bottom gradient through the narrows or along the bend, which will permit more rapid flow with less surface adjustment, so that the backwater effects may diminish. The course of Arroyo Seco Canyon rapidly alternates between straight stretches and abrupt bends, as well as relatively broad and narrow portions. At each change of direction or width opportunity arises for backwater effects. Debris cones from the tributaries, partially blocking the canyon, also have the effect of constrictions.

The debris cones built out from the tributaries are probably due mainly to the change of gradient between tributary and main canyon. Some of the small tributaries from the west (such as Agua Canyon) have slopes of 500 feet to the half mile, which is more than four times the average slope along the main canyon. Many of these short tributaries had road fills in their upper portions. During the flood this material was delivered to the tributary mouth where much of it came to rest. The larger volumes of flood water in the main canyon, already taxed with heavy loads, were unable to remove the material as it accumulated.

An example of deposition controlled by a constriction and a bend was given in connection with site 9 (Pl. 6, fig. 2). An example of the constriction caused by a debris cone from a tributary was given in connection with Brown Canyon (Fig. 8). In the light of the analysis (Fig. 18), backwater effects would have occurred upstream of Brown Canyon, with deposition in the zone of deepened water. After the surface slope was adjusted (combined with some adjustment of bed slope) the remaining load could be carried through with no deposition until the constriction was passed. Conditions upstream from Brown

Canyon substantiate this analysis, inasmuch as the main canyon there was so choked with debris that it had to be cleared to allow space for the stream and temporary road.

Log and boulder jams may arise in several manners. If logs are swung sidewise even in a broad part of the valley, the abrupt locking which results creates in effect a dam, with its immediate upstream response,—a slackening of velocity and a piling up of water. The partially cleared water pouring over the jam or around it can produce scour downstream. The coincidence of jams with bends and constrictions is apparently due to the much greater chance of logs or boulders locking. The net result is a complex deposit involving the elements of damming and constrictions. A hydraulic jump would favor the formation of jams by its abrupt slackening of velocity in the transition zone.

The presence of reverse size gradients in the jams is still not adequately explained. The only work bearing on this subject which has come to the writer's attention is an experiment by Straub (1940), in which natural river gravel was studied in a flume. When the load was added slightly faster than the current could carry it, the bottom slope was built up, and the coarsest particles were carried to the downstream end of the flume. Mechanical analysis of samples taken at intervals along the flume showed an increase in mean size from 2.5 mm. to 5.0 mm. over 24 feet. The original sediment had a mean size of 4.0 mm. Thus the finer material was left near the upstream end, and progressively coarser debris moved down the flume. According to Straub (1940, p. 183):

> "Depending upon the flow conditions and type of material, segregation upon deposition may take place either from fine to coarse (from one end of the channel to the other) or the reverse. Much additional study is required both in field and laboratory before any definite conclusions can be drawn as to the laws governing this occurrence."

Although the physical law underlying this situation is not known, apparently the controlling conditions were a thickening of the bed in a downstream direction as deposition progressed. If this experimental work may be applied to the present instance, very likely the lenslike deposits of the jams furnished the necessary conditions by building up most rapidly at the downstream end, thus decreasing the gradient and setting the stage for this type of size sorting.

The response of the sediments in terms of size and shape along the canyon as a whole conformed strikingly to the mechanics of their movement and deposition. The observations in Arroyo Seco indicated that in comparable sections of the deposits the average size and degree

of sorting showed no systematic variations along the valley. Similarly, when the data are confined to one rock type, as the Lowe granodiorite, the average sphericity varies nonsystematically, whereas the roundness increases downstream. Inasmuch as size, shape, and density largely control the settling velocity, these attributes should reflect the dynamical conditions at the site of deposition. Roundness,—of secondary importance in settling velocity, but of primary importance as an index of abrasion—is a function of the wear on the pebble over the distance traveled. These same relations were noted in San Gabriel Canyon, from which it was inferred that essentially similar dynamical conditions were present at successive log jams; that is, the abruptness and order of magnitude of the velocity decrease was the same, because the size and shape response was the same.

In Arroyo Seco (and perhaps in San Gabriel Canyon) the samples include both jam material and alluvial plain deposits. The dynamic effect of constrictions and sharp bends on large masses of water moving moderately rapidly may be considerable, and responses in velocity adjustment may be rapid, especially if hydraulic jumps are formed. The frequent coincidence of log jams and bends argues that many of the alluvial plain deposits upstream of the bend may have formed under conditions nearly as vigorous as in the jams. Where constrictions cause deposition, the response of the sediments upstream is strictly dependent upon conditions at the constriction, because of the control of the water profile surface, and hence of the velocity, by the constriction itself. Therefore it may often be necessary in studying sediments to locate the control point in order to evaluate the results.

PARTICLE WEAR AND SELECTIVE TRANSPORTATION

The movement of debris along natural streams is accompanied by two basic processes going on simultaneously: selective transportation and particle wear. Selective transportation is reflected in the segregation of certain materials into bed and suspended load, with the simultaneous formation of a lag deposit. Selection proceeds largely on a basis of size, shape, and density. Highly spherical particles, for example, may outstrip less spherical ones in bed-load movement, whereas the less spherical particles outstrip more spherical ones in suspension transport. Wadell (1934) has discussed these effects in connection with his concept of sphericity. Simultaneously with movement of the particles, especially in bed-load transportation, size, shape, and roundness change as a result of particles rubbing or striking against each other. Marshall (1927) has shown that these effects may involve impact (breakage), abrasion (rubbing of particle against particle), or grinding (crushing of

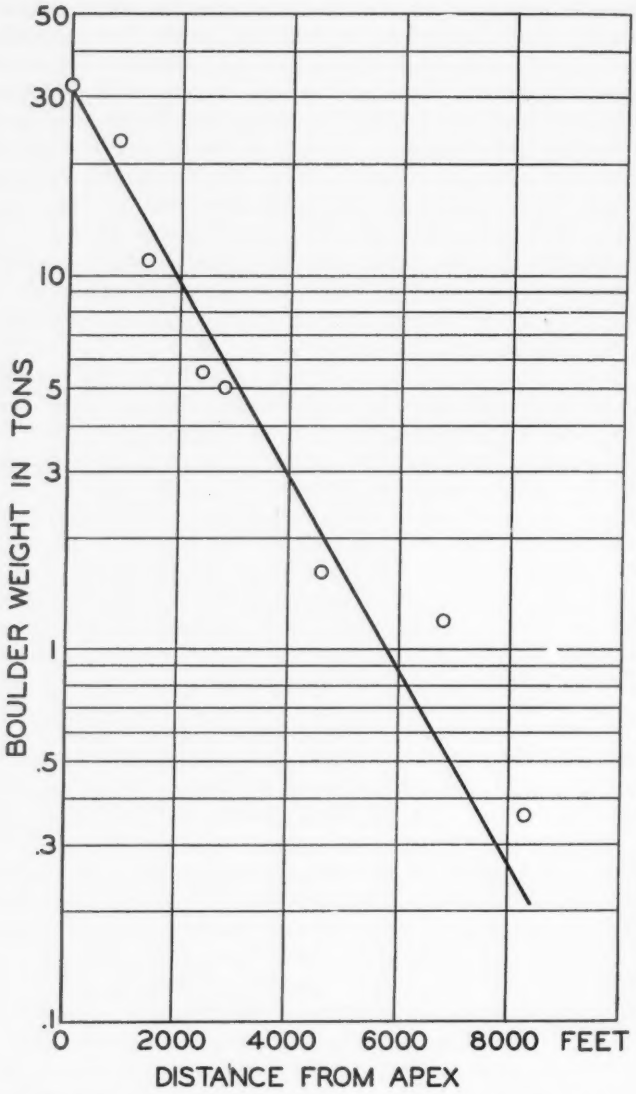

FIGURE 19.—*Variation in boulder size caused by La Crescenta flood, 1934*

(Data from Chawner, 1934.)

small particles by large ones), all included in the term "wear". Which effect predominates depends upon the rigor of the process and the nature of the rocks. In general the average size of the particles decreases downstream over any considerable stretch. This is frequently attributed to wear, but it could also be due to lessened competence of the stream caused by a continually decreasing gradient. It is not possible, apparently, to segregate the two effects merely from the observed reduction in mean size; the problem is one of two unknowns and a single equation.

Under conditions of laboratory abrasion the size and shape change much less than the roundness (Krumbein, 1941b), so that considerable rounding may occur with negligible size and sphericity effects. In streams where abrasion is the only process of wear, apparently any rapid size change should be attributed to selective transport. Impact breakage has not been extensively studied, but the process undoubtedly results in a rapid decrease in size compared to abrasion alone. In steep canyons where greater rigor may permit impact breakage, the observed size decrease may actually be due to wear.

The relative effects of the two processes can be evaluated in the laboratory. Particle wear studies, in which the composition of the load remains constant, shed light on processes of abrasion and breakage independent of selective transport. Flume studies on the other hand permit an evaluation of selection where particle wear is entirely negligible. Rarely does only one effect occur in nature. The reversal of the size gradient in jam deposits can only be attributed to selective transport. Chawner (1934; 1935) measured a number of large boulders carried onto the alluvial fan during the La Crescenta flood of 1934. He tabulated the size of boulders in terms of their distance from the apex of the cone (Fig. 19). Despite the few observations, a marked size decrease is apparent over a short distance. The straight line indicates an exponential trend. The half-distance of the boulder size is found by determining the distance at which an ordinate of the straight line is half the value at its origin. This distance is 1100 feet from the canyon mouth. That is, boulder size decreases by half in the first 1100 feet across the fan. Here the short distance of transport, combined with the decrease in water velocity attendant upon its spread over the fan, indicates control by selective transportation.

Along Arroyo Seco the Lowe granodiorite boulders decrease markedly in size downstream. This effect is attributed to a statistical average of conditions over a long period of time, rather than just to one flood. Considering all rock types there was no over-all size gradient among the boulders along the canyon, due very likely to the continual influx

of new material and to essentially similar conditions in the jams and other deposits. The only clue to possible effects of wear and selective transportation came from the data of Figure 11. The break in the straight line, which occurred near the zone between erosion and aggradation in the canyon, may reflect a change in the conditions of wear (as from breakage to abrasion); it may reflect an increase in the effects of selective transportation; or it may indicate a relative adjustment in both processes. The magnitude of the change is indicated by a ratio of the half distances of the two curves. In the steeper curve the half distance is 3.7 miles, and in the gentler curve it is 10.0 miles. Thus, on the average, the rate of apparent size decrease in the upper part of the canyon is more than 2.5 times that in the lower reaches.

In more detailed studies of wear and selection it may be necessary to include shape and roundness of the boulders. Angularity of the boulders over any considerable stretch may indicate continuous impact breakage, whereas progressive rounding may indicate less rapid abrasion. No detailed measurements were made, but angular boulders with evidence of fresh fracture were rare. It is the writer's present opinion that selective transport is on the whole more important than wear even during floods, especially in the lower part of the canyon.

SUMMARY AND CONCLUSIONS

The present study extends and amplifies the previous year's work on flood deposits. Certain tentative conclusions reached in San Gabriel Canyon were strengthened by evidence from Arroyo Seco; other inferences were modified by additional data. The combined studies furnish sufficient data for at least a qualitative reconstruction of the physical processes which operated during the flood.

Field work showed three main kinds of deposits formed by the flood: log and boulder jams at bends and constrictions; typical alluvial plain deposits in broad parts of the canyon; and alluvial cones at tributary mouths. These deposits were controlled by one or more of the following factors: abrupt decrease of gradient as at tributary mouths; abrupt velocity decrease associated with locking logs or boulders; velocity decrease caused by water profile adjustments at constrictions; and perhaps abrupt velocity decrease caused by hydraulic jumps.

Laboratory analysis of gravel samples showed no systematic variation in size or shape from corresponding parts of successive flood deposits, although in any single deposit (mainly jams) the size increased markedly in a downstream direction. Pebble roundness showed a systematic increase downstream along the entire valley. Pebble orien-

tation showed uniform upstream imbrication even in the steep upper part of the canyon, and further studies in San Gabriel Canyon showed that reverse imbrication found there was due to local causes.

A limited amount of data were presented on the relative effects of boulder wear and selective transportation along the canyon. These data suggest a more rapid apparent size reduction in the upper reaches of the canyon, but they do not permit an evaluation of the direct cause of the size decrease.

It became increasingly evident during the study that the basic unit for sedimentary investigations is not the sample, but the entire depositional environment. The analysis of sediments should be integrated with physical principles of particle transportation and wear, but such integration is at present hindered by lack of complete data on flood discharge, sediment load, and velocity changes along the canyon. The superposition of large flood phenomena on more normal processes of stream work also introduces a complicating factor. It is believed, however, that the geological approach offers the most direct means of reconstructing flood conditions until such time as more direct measurements of flood phenomena are available.

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